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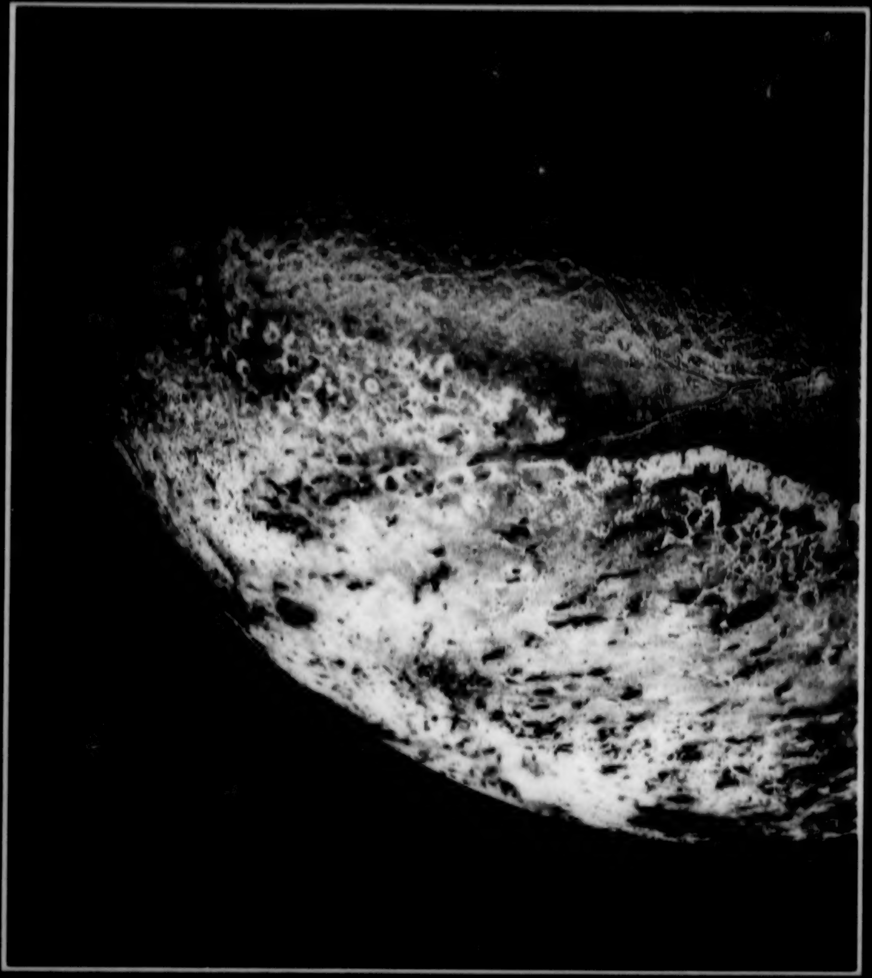

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Preface

During the past three decades we have completed the first phase of the exploration of all major bodies (with the exception of pluto) in our Solar System. Two recent flight mission successes, the flyby of Neptune by Voyager and the mapping of Venus by Magellan, have provided exciting new discoveries and key new information about these planets. These new data are already providing information that will increase our understanding of both planets as well as other information that can be used for planetary comparison studies.

Discoveries made as a result of exploration of the Solar System have had an enormous effect on our understanding of the planets, including Earth, and the way that we regard our place in the universe. This has been and will continue to be one of the most exciting and scientifically important eras in human history—the era of Solar System Exploration.

NASA's Planetary Geosciences Programs (the Planetary Geology and Geophysics and the Planetary Material and Geochemistry Programs) provide support and an organizational framework for scientific research on solid bodies of the solar system. These research and analysis programs support scientific research aimed at increasing our understanding of the physical, chemical and dynamic nature of the solid bodies of the solar system: the Moon, the terrestrial planets, the satellites of the outer planets, the rings, the asteroids, and the comets. This research is conducted using a variety of methods: laboratory experiments, theoretical approaches, data analysis, and Earth analog techniques. Through research supported by these programs, we are expanding our understanding of the origin and evolution of the Solar System.

This document is intended to provide an overview of the more significant scientific findings and discoveries made this year by scientists supported by the Planetary Geosciences Program. To a large degree, these results and discoveries are the measure of success of the programs.

The scientists responsible for these major scientific discoveries and advances have provided short articles describing their research and its importance in the context of the larger problems in Planetary Geosciences. These short articles have been edited and incorporated into this document by an editorial board that consists of Maria Zuber (NASA/Goddard Space Flight Center), Odette James (U. S. Geological Survey), Glenn McPherson (Smithsonian Museum of Natural History), Jonathan Lunine (University of Arizona), and Roger Phillips (Southern Methodist University).

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ANCIENT MARTIAN OCEANS

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Discoveries in science include both new information about nature, and new ways of interpreting the scientific information already in hand. We report here on a new understanding of global environmental change on Mars achieved from a study of existing data. This view of climatic change on Mars struck us as we pondered a great variety of perplexing surface features on the planet.

For many years, like other planetary geologists, we were puzzled by numerous enigmas on Mars. Why did some portions of the surface seem to have very low erosion rates, while other areas have forms indicating intense degradation by erosional processes? Why did numerous valleys and channels form during the martian past, when the modern climate is too cold and dry for active water flow? How could water have moved to replenish ancient streams when the modern atmosphere has less than 1/100 the pressure of that on Earth?

The answer to these and other martian dilemmas can be provided by past episodes of temporary ocean formation. We conclude that vast areas of the northern lowland plains on this dry, desert planet were sporadically inundated by huge quantities of water (Fig. 1). Repeated formation and dissipation of this ocean, which we name Oceanus Borealis, resulted in relatively warm, wet climatic epochs that favored the development of glaciers in the southern hemisphere and highlands of Mars. Lake formation, melting of permafrost, active landsliding, and erosion of old craters all seem to have occurred during temporary warm, wet (maritime) intervals of martian climate. When the oceans gradually evaporated or froze, the planet returned to its cold, dry condition with its water trapped as ground ice in underground permafrost. It is this cold, dry mode that presently characterizes Mars.

Earliest martian geological processes include very high rates of impact cratering and widespread volcanism. During this early epoch (Noachian time), rainfall and surface-water flow in valleys occurred, probably because of a persistent hydrological cycle involving long-term residence of water in an ocean. As impacting rates declined and volcanism became concentrated in local areas, this primordial ocean

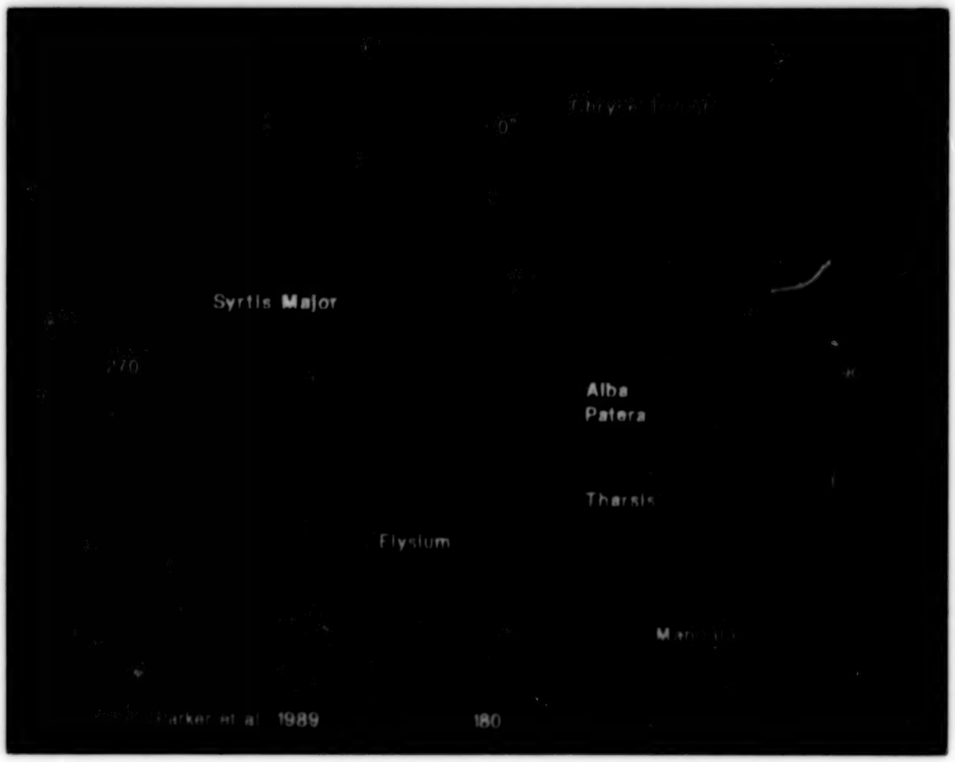

Fig. 1. Polar projection of shoreline features in the northern plains of Mars as interpreted by Timothy J. Parker, R. Stephen Saunders, and Dale M. Schneeberger. Directions of outflow channel inputs to Oceanus Borealis are indicated by the blue arrows. A possible high stand of Oceanus Borealis is indicated by the green boundary line and blue shading. The red boundary line indicates the extent of a smaller, younger water body. The exact ages of these episodes will remain somewhat speculative until further work is done. The oldest phase seems to have occurred after the cessation of high impacting rates (in Hesperian time). The youngest ocean formation may have occurred during Amazonian time, the youngest period of martian history (but still as much as a billion or more years ago).

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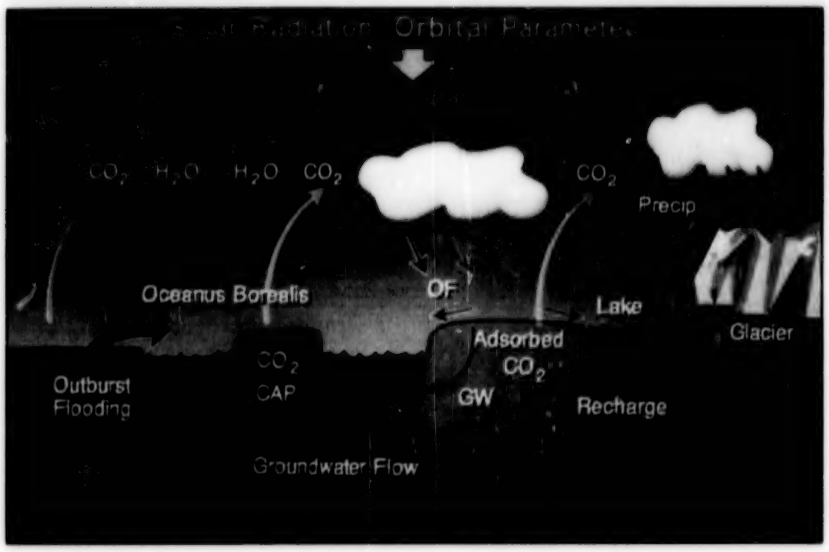

Fig. 2. Schematic representation of various ocean-land-atmosphere interactions associated with the formation of Oceanus Borealis. Volcanism and heat flow (red arrows) drive hydrothermal groundwater circulation and outburst flooding from permafrost-ground ice storage reservoirs in the upper martian crust. Outburst flooding (large blue arrow) delivers immense discharges to northern plains, resulting in large fluxes of water (from evaporation) and carbon dioxide (from water attack on the polar CO_2 cap and from solution in released ground water). Warmed by these greenhouse gases, the maritime atmosphere precipitates snow to form glaciers in the southern polar regions to supply lakes and overland flow (OF) features (stream valleys) at lower latitudes. Adsorbed CO_2 may be released from the upper layers of the martian crust to further enhance the greenhouse warming. Recharge and ground-water (GW) flow will complete the cycling of water back to its major planetary sink at Oceanus Borealis. Unlike Earth, however, this dynamical cycling did not function continuously in martian history. The ocean or smaller lacustrine bodies formed sporadically.

disappeared. Its sporadic reappearance in later martian history is recorded by Oceanus Borealis sediments and landforms that were described by several investigators prior to our work. The oceanic evidence was assembled most impressively by T.J. Parker, R. S. Saunders, and D. M. Schneeberger of the Jet Propulsion Laboratory, California Institute of Technology. Stimulated by their work and by a vast array of studies by many planetary geologists, we developed the following global model of ocean formation on Mars.

Our model arose intuitively from experience with martian phenomena and from hypothesizing the origin and consequences of an ocean. Figure 2 shows the major flows of mass and energy associated with this martian hydrological cycling. During later martian history, huge concentrations of molten rock (magma) were concentrated at one local region of the planet, the Tharsis volcano area. Massive and rapid emplacement of magma beneath this bulging zone of anomalous heating melted large amounts of ground ice, driving it into fractures on the margins of the Tharsis bulge. The water burst on to the surface at great outflow channels heading at these fractures. Driven by volcanic heat, the cataclysmic outburst floods of water and mud carved immense, spectacular channels, tens of miles wide. The warm water inundated the northern plains of Mars, vaporizing the north polar cap of carbon dioxide ice.

The cataclysmic ocean formation had an immediate climatic influence. Both water (evaporated from the sea) and carbon dioxide (from the polar cap) are greenhouse gases. Just as human burning of fossil fuels is causing a global warming of Earth, so the martian floods induced a cataclysmic warming. As martian temperatures rose, other water, frozen in upland permafrost, was released to flow into lakes or the Oceanus Borealis. The climate moved to its maritime state, with precipitation possible.

During the late phase of ocean formation, much of the precipitation fell as snow. Particularly near the south pole and in upland areas, the snow accumulated to thicknesses sufficient to form ice. As the snow and ice built up, it flowed as glaciers. The geological evidence indicates that the glaciers advanced and retreated, much the same as occurred in the Ice Ages of Earth. Because the late Mars ocean was relatively short-lived, this martian glacial epoch was also very brief. Figure 3 shows the planetary relationships during transient ocean formation.

Oceanus Borealis was probably a persistent feature early in Mars history. At that time the planet was experiencing a relatively high rate of impacting objects. The dense water and carbon dioxide atmosphere allowed precipitation as rain, resulting in the widespread valley networks of the martian uplands. However, water was being lost because of dissociation in the upper atmosphere of the planet. The hydrogen was lost to space while the oxygen contributed to the red color of the planet by oxidizing various materials. Eventually water loss and precipitation of the carbon dioxide as carbonate rock reduced the atmospheric pressure below the level for maintaining the ocean. Most of the water was sequestered into the very permeable rocks of the planet, where it comprised ice in a permafrost.

The ocean was able to reform much later in the planet's history because of the Tharsis volcanism described above. Oceans formed by cataclysmic water outbursts were smaller than the original one because of the water loss by hydrogen

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escape. However, the younger oceans were big enough to temporarily modify the climate, producing the enigmas that had bothered us about the martian surface.

The above model has completely changed how we view Mars. Instead of a cold, dry world with strange features inconsistent with its modern climate, we now see a more Earth-like planet. During its early history, Mars was very water-rich, and it seems to have had a watery atmosphere. Unlike Earth, the rocks and landscapes of Mars' first billion years of history are well preserved. Did life form in the early martian ocean? If so, fossils of that life might be present and available for discovery by a future geological exploratory mission. On Earth, we have little or no material left from that early period because of the active reworking of the planet's surface. The stable martian surface is a unique record of the same watery past that characterized our own planet.

Mars and Earth are the only known planets on which water has moved in a dynamic cycle from its reservoir in the ocean to rainfall, flowing rivers, and back to the sea. This same cycle, the hydrological cycle, is the source and the maintainer of life on Earth. Might it have served as the source for life on Mars as well? We may find the answer to the origin of life on Mars rather than on Earth.

Sometimes science is viewed as a monolithic enterprise of computers, laboratory equipment, and individual theorists. We forget that science is a group exercise in which people make sense, a kind of common sense, out of the world in which they live.

We feel we have found a way to make sense out of what seemed to be perplexing problems of past environmental change on the planet Mars. The way that this has changed our view gives hope that a similar sense can be achieved about the perplexing problems of future environmental change on the planet Earth. Is there something in the existing information that we are just not seeing? Can one simple idea give us the clue to how it all fits into a pattern?

We have a new confidence on how Mars works as a planet, how its water-related systems have evolved through time. We need a similar confidence for Earth. Rather than idealized future "scenarios" given to us by computers, we need an understanding of how the whole planet works. If we can figure it out for a slightly smaller, slightly colder version of Earth known as Mars, that process of common sense should allow us the same revelation about Earth and its global changes. As the philosopher William Clifford once said of science "...the truth at which it arrives is not that which we can ideally contemplate without error, but that which we may act upon without fear."

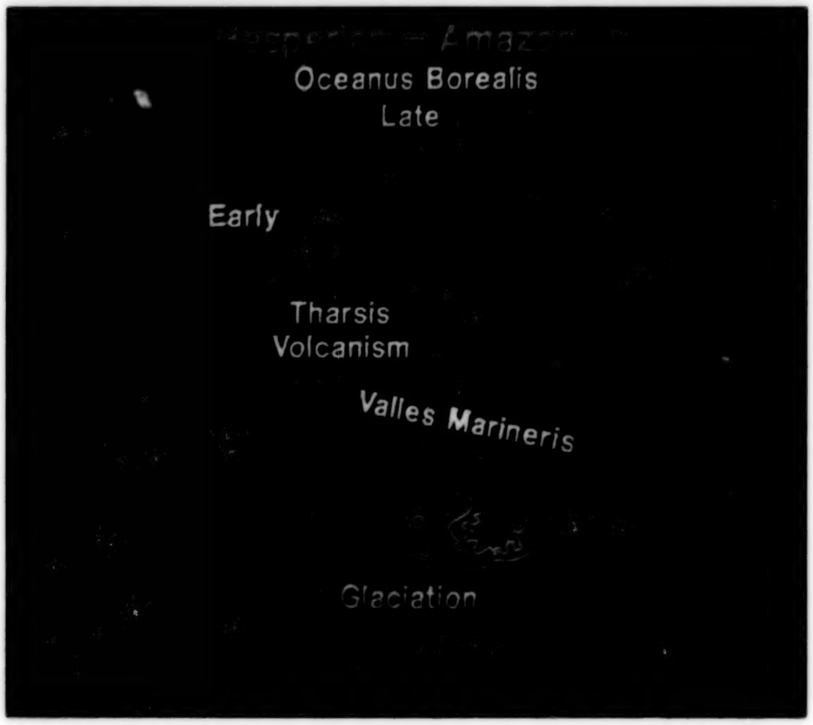

Fig. 3. Sketch map of the western hemisphere of Mars showing the relationships between areas of glaciation in the southern hemisphere and Oceanus Borealis in the northern hemisphere. The outflow channel discharges were directed as shown by the blue arrows. Note the relationship of the discharge sources, including Valles Marineris, to the zone of Tharsis volcanism. Our preliminary calculations indicate that the immense discharges of the outflow channels could have filled the Oceanus Borealis basin in a period between about 2 and 40 weeks. The indicated water volumes could have been supplied from a water-rich permafrost layer on the order of 1 kilometer thick that was drained to the outflow channel heads with energy supplied by massive volcanism. It is likely that volcanism concentrated at Tharsis because of a zone of unusually high heat flow that produced sporadic upwelling of molten rock from the deep martian interior.

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CONSTRAINTS ON THE STRUCTURE AND EVOLUTION OF THE THARSIS REGION OF MARS

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The immense Tharsis rise of Mars is of a scale unrivaled in the solar system. It is a roughly circular area 2000 miles in diameter that ascends more than 7 miles above the surrounding plains (Fig. 1). Dominating the western hemisphere of the planet, it is intensely fractured by grabens (narrow, fault-bounded valleys) that extend radially from its center to hundreds of miles beyond the rise and is ringed by compressional ridges that formed over 1200 miles from its center. Its immense geological influence, involving an entire hemisphere of Mars, gives it a central role in the thermal and tectonic evolution of the planet, and has stimulated a number of studies attempting to determine the structure beneath the surface of this unique feature and the sequence of events responsible for its formation.

Both geophysics and geology can provide clues to these problems. Geophysical evidence is obtained primarily from the ratio of gravitational acceleration to topography as a function of wavelength. Most large-scale elevated topography on Earth is found to be in "isostatic compensation." This means that large mountains are virtually "floating" on the relatively ductile upper mantle, which requires that the crust beneath them "compensate" for the extra mass of the mountains either by a greater thickness or lower density. The depth at which this compensation occurs, usually a few tens of miles in the Earth, can be derived from the gravity-to-topography ratio.

For Tharsis, the calculated depth of compensation derived from these ratios ranges from 650 miles at a wavelength of 6600 miles to about 60 miles for the shorter wavelengths. It can be shown that this implies a large excess of gravity at long wavelengths relative to that which can be explained by simple compensation models. There are a limited number of ways to realistically accommodate this. One method is to assume two-level isostatic compensation. This model requires massive removal of crust (decreasing its thickness by as much as 40 miles) and support by an immobile 200 to 300 mile thick layer of low density mantle that has undergone some degree of melting, with the melt products (magma) having been moved to the surface by volcanism.

Another possibility is that the huge topographic load of Tharsis is supported by mantle convection, which would require a single, immense, upwelling plume that has not moved significantly with respect to Tharsis for more than a billion years. The third possibility is that the load is largely supported by the strength of the lithosphere (the rigid outer shell of a planet), just as a beam supports the structure of a building. In this case, a modestly thickened crust is required with a lithosphere at least 60 miles thick, capable of supporting large elastic stresses by bending.

Geology provides information on strain in the crust as a function of time. Faulting activity around Tharsis has been both long-lived and varied, involving local, regional, and full hemispheric events. Three distinct hemisphere-wide episodes have been recognized. A radial system of grabens centered on Syria Planum marks the first unequivocal tectonic event in the formation of Tharsis. This fracture system can be traced from near the center of the rise to well beyond its flanks. Somewhat later, massive outpourings of fissure volcanics formed the surface of Lunae Planum and similarly aged plains surrounding Tharsis (the so-called ridged plains). Shortly after deposition, these plains were deformed by compressional ridges with a pattern concentric to a point in Syria Planum. Still later, another major radially oriented normal faulting event occurred, this time centered near Pavonis Mons.

The geological and geophysical information described above can be linked via theoretical stress models. In broad terms, three distinct fault patterns can be generated at the surface of a spherical shell (e.g., the lithosphere) that supports a roughly circular topographic high whose dimensions are comparable to the radius of the planet. If the topography is due to uplift forces from beneath and the shell bends elastically, circumferentially oriented extensional features (e.g., grabens) within the raised area and radially oriented compressional features

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(e.g., thrust faults which may be expressed as ridges at the surface) in the periphery are predicted. In contrast, a surface load that causes subsidence will induce circumferential ridges within the loaded region and radial extensional faulting farther out. Isostatic support of topography without support by bending stresses will result in radially oriented normal faulting in a region around the topographic center and circumferential ridge formation near its edges. Stresses due to convective support of topography are still under study, but the deformation patterns are likely to be similar to those of the uplift case.

These results place strong constraints on the evolution of Tharsis. Uplift will form circumferentially oriented grabens, which are not common on Tharsis. Thus, this mechanism can be ruled out for all but perhaps the earliest stages of formation. Circumferentially oriented ridges near the edge of the rise are a feature of both isostasy and subsidence, with the former extending the ridge zone outward and the latter extending it inward. The observed distribution of ridges tends to marginally favor an isostatic regime. The most clear-cut distinction between these two cases, however, is in the respective regions of radial extensional faulting. Inner radial faulting is possible only for isostasy, whereas outer radial faulting can occur only with subsidence. The conditions under which these two patterns form should be mutually exclusive. However, detailed stratigraphic mapping has shown that both inner and outer radial faulting occurred contemporaneously over a considerable interval of Tharsis history. Thus, there appears to be a fundamental contradiction between model results and observation.

The gravity observations, along with the difficulties inherent in creating a realistic isostatic structure, argue for flexural support of the Tharsis load. Additionally, most of the regional tectonic deformation, with the notable exception of the inner radial faulting, is consistent with subsidence of the lithosphere. We contend that the formation of the inner radial fault pattern is a natural consequence of the thickened crust and higher heat flow resulting from the extrusive and intrusive volcanic construction of the Tharsis rise.

The stress models described above assume that the load is supported by a homogeneous elastic shell. But the actual mechanical behavior of the lithosphere is undoubtedly more complex. Laboratory

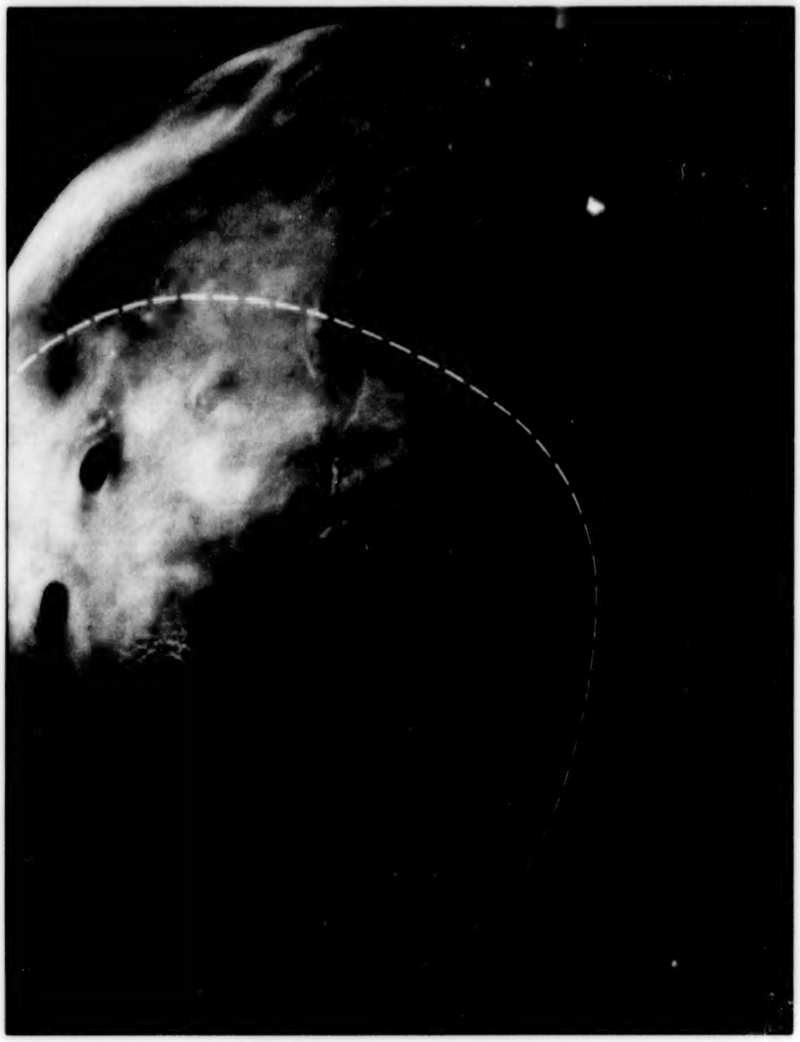

Fig. 1. Photomosaic of western equatorial region of Mars, displaying the vast extent of Tharsis (the dashed line approximates the three-mile elevation contour). Shown prominently are Valles Marineris (the large trough extending about 2000 miles in the center of the image) and the three large Tharsis Montes shield volcanoes (each more than 16 miles in elevation). The center of the Tharsis rise is within the jumbled series of troughs called Noctis Labyrinthus at the western end of Valles Marineris. The elevation increases by about 7 miles from the eastern end of Valles Marineris to Noctis Labyrinthus. Concentric compressional tectonic features and radial extensional tectonic features (Valles Marineris is a huge rift) cover most of the field of view and are centered near the high point in Noctis Labyrinthus. (Image courtesy of Al McEwen, U.S.G.S., Flagstaff.)

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and field measurements on Earth indicate that the lithosphere may have two strong zones separated by a weak ductile layer in the lower crust. The existence of this layer depends on the crustal thickness and thermal gradient, with larger values of either of these parameters making such a layer more pronounced.

If Tharsis was formed by construction, it will act as a flexural load on the elastic lithosphere. Far from Tharsis, the elastic lithosphere will include both the crust and upper mantle. However, within Tharsis itself the thickened crust and high heat flow will act to decouple the upper crust from the strong zone in the upper mantle. In this situation, the upper mantle strong layer, which constitutes most of the lithosphere in either case, will deform as part of the global shell, transferring flexural stresses and displacements to the rest of the shell. The relatively thin, brittle upper crustal layer will deform not as part of the greater shell, but rather as a spherical cap with a lubricated lower surface and a peripheral boundary which is fixed to the global shell (Fig. 2). This cap will respond primarily to spreading forces and membrane stresses induced by the subsidence of the lithosphere. Both of these processes will lead to radial faulting within the highland area. Outside this region of decoupled crust, faulting will be due to flexural stresses caused by the overall lithosphere subsidence. Radial compression is likely to be concentrated near the boundary of the two regions, resulting in the enhanced formation of concentric wrinkle ridges.

Thus, this model provides a self-consistent explanation for the key geophysical and geological observations relating to the structure and history of Tharsis. If confirmed, it can provide important constraints on the crustal volume, volcanic history, and thermal evolution of Mars. The Mars Observer mission, due to be launched in 1992, will provide detailed gravity and topography maps of the planet. These new data will provide important information to test this and other competing models for the formation and structure of Tharsis.

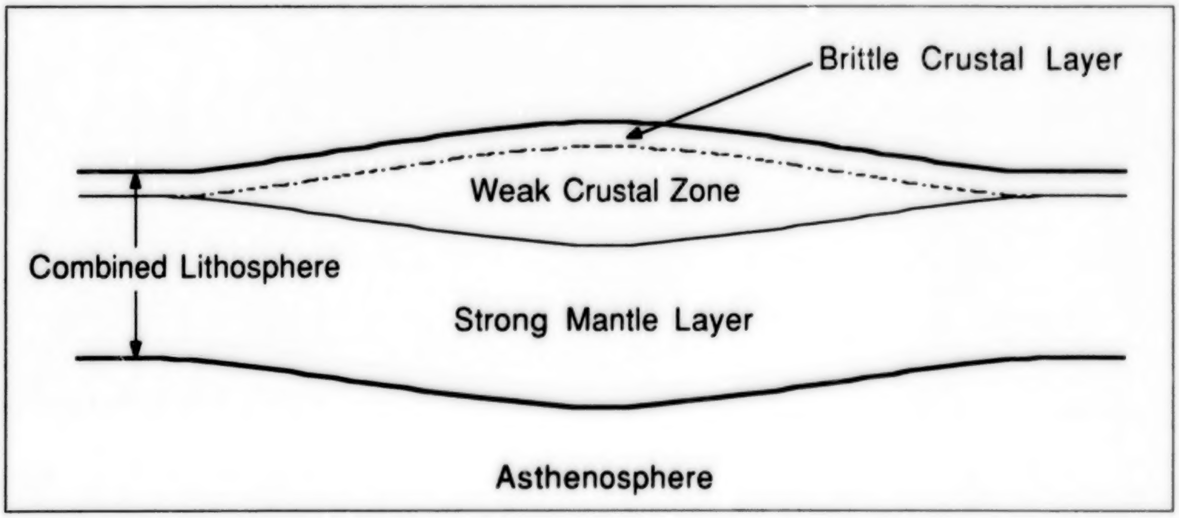

Fig. 2. Schematic diagram of the proposed model, showing the thickened crust beneath the Tharsis bulge which results in a thin, brittle, elastic crustal layer at the surface that is separated from the strong elastic layer in the upper mantle by a weak, ductile zone in the lower crust.

"NIGHTFALL" AND THE ORIGIN OF PLANETARY SYSTEMS

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One of the better known works of Isaac Asimov is entitled "Nightfall." It deals with a civilization on a planet that orbits in a sextuple stellar system (i.e. six "suns"). A consequence of this unusual celestial configuration is that the planet experiences daylight for some two millennia; nightfall is a rare and destabilizing social experience for the denizens of this fictitious planet.

Although sextuple star systems such as that in Asimov's story are rare, nature does appear to preferentially form stars in at least binary (two star) systems. It is estimated (from limited observations) that perhaps seventy percent of stars are members of binaries, and the percentage could be significantly greater. All of the nearest ten star systems to the Sun are multiple; the closest star of all, α-Centauri, is a triple star system. In fact, it is stars such as our Sun, which is a single star, that appear to be unusual.

Given the current interest in the possible existence of life elsewhere in the universe, it is natural to ask the related question of whether there are planets around other stars and, if so, what kinds of stars or star systems are most likely to have them. We do not yet have the means to detect directly planet-sized bodies around even the nearest stars, so observations cannot yet answer the question of whether multiple star systems can have planets. Unfortunately, the existence of planetary systems in association with multiple stars is also a very difficult theoretical question. In the first place we don't know whether the process by which multiple stars are formed permits formation of planetary systems, but even if it does there is the difficult physics problem of whether such planetary systems could remain dynamically stable under the gravitational influence of two or more suns.

The problem of stability of celestial systems is nearly as old as astronomy itself. For systems with more than two components (e.g., two stars and one or more planets), a reasonable starting place for calculating stability is the classic so-called "three-body" approximation. The work described below is from research that we and others have conducted over the past decade, as parts of beginning efforts to examine theoretically how and under what conditions stable planetary systems can evolve, and whether analogs to Asimov's world, or to the world depicted so vividly in the painting by astromer/artist William Hartmann of Arizona (Fig. 1), exist in nature.

It is convenient to discuss "three-body" planetary models in terms of three possible configurations between a planet and its associated multiple stars. In one, referred to as the "outer planet" case, the small planetary body revolves around the binary stellar system as a whole rather than around one of the stars. In the second, called the "inner planet" case, the planet revolves around the more massive of the two stars. The third is referred to as the "satellite" case, in which the planet revolves around the less massive of the two stars (similar to the way the Moon revolves around the Earth in the Sun-Earth "binary system").

A recent attempt to analyze a three body celestial system was by F. Graziani (University of Minnesota) and D.C. Black in 1982, who studied the very simple case of two equal sized planets orbiting a single star. Using a relatively simple mathematical expression for the relative distances and masses of the three bodies, they showed that a critical mass combination led to the onset of instability. Their expression was generalized later by one of us (Black) to describe the onset of instability in all kinds of planetary systems, but especially one involving multiple stars (still assuming just three bodies).

Independently, V. Szebehely (University of Texas) and his co-workers used a very different approach to examine the stability of three-body systems. They were able to show that the three classes described above give rise to three distinct stability curves. What is significant is that when the results of Szebehely and his co-workers are cast in terms of the same parameters used by Graziani and Black, they match well the results of the Graziani-Black model. Furthermore, the Graziani-Black

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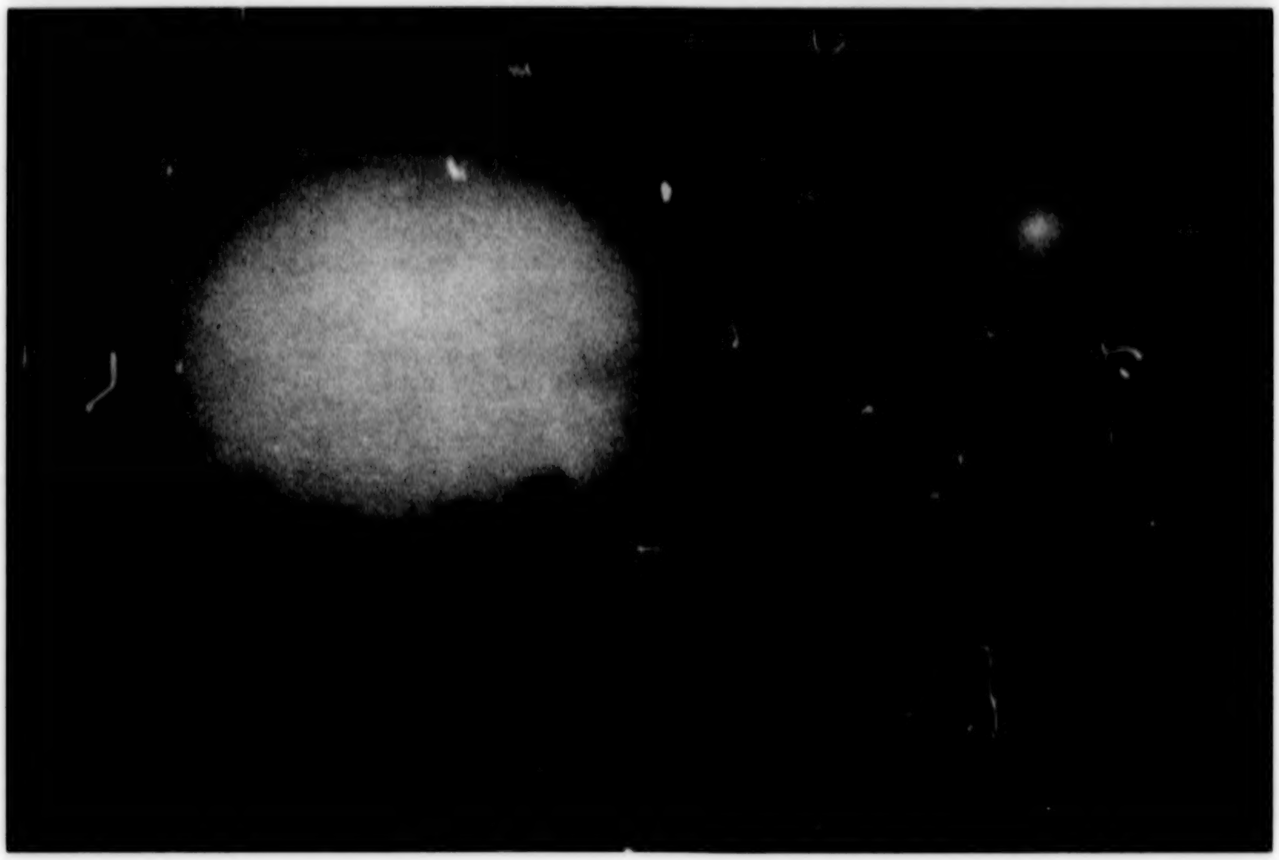

Fig. 1. A view of the binary star system comprised of the star Antares and its white dwarf companion from the surface of a hypothetical planetary companion to those stars (painting by William Hartmann).

model appears to describe stability for both the restricted and the general three-body problem. The agreement between the two approaches suggests that the approximation of Black and Graziani is a reasonable description of the onset of instability in a three body system.

Planets And Binary Systems

While these studies do not yield a single parameter that describes whether planetary systems associated with binary stellar systems will be stable, they do provide a means of placing limits on the sizes and placement of orbits for stability given the masses both of the stars and of the hypothetical planet. Observations indicate that the most likely outcome of binary star formation is that the stars will be of nearly equal mass. For such a case, the calculations suggest that stable planetary orbits are possible if either (1) the planet revolves outside the binary at a distance that is at least five times the separation of the binaries, or (2) the planet revolves about either of the stars at a distance that is no greater than about one-fifth the separation of the binary stars.

The studies cited here indicate that planetary systems can indeed be stable in association with binary star systems, but there are restrictions on where the planets can be. What these studies do not address is whether planets would ever form in binary star systems in the first place. That is a far more difficult problem, one whose solution will follow from a better understanding of how planetary systems form in association with single stars, such as the Sun. To that problem we now turn in the final part of this paper.

Stability And The Formation Of Planetary Systems

Our understanding of how the solar system formed has increased significantly in the past two decades. One key to this increase has been the development of more complete theoretical models of what are thought to have been the major processes in the formation and early evolution of the solar system.

It is generally accepted that the planetary members of the solar system formed from a rotating disk of gas and dust. Observations of young stars indicate the presence of such disks, supporting the notion that disks are a basic result of the star formation process. It is thus not unreasonable to suppose that many stars have planetary companions even though, as noted earlier, none have been detected at this time (popular reports to the contrary). There is as yet little understanding of what controls the masses and locations of the planets that might form out of these disks. Our admittedly simplistic approach to this last question is to examine how orbital stability considerations of the type described above might set limits on forming planetary systems.

We start by using the mathematical model for a presolar gas disk as proposed by T. Nakano of Tokyo University, and apply it to the case of forming giant gaseous planets like Jupiter and Saturn. We necessarily make certain assumptions, for example, that all of the mass in a given part of the disk is accreted into a planet. Our goal is to determine what kinds of disk configurations (size, density, mass distribution) permit the formation of pairs of planets, having masses comparable to those of Jupiter or Saturn, that are orbitally stable.

Results from a series of calculations indicate that pairs of very large planets (Jupiter-like) should be able to form in stable orbits around stars with masses less than 3-4 times the mass of the Sun, provided the nebula from which they form is of modest density. However, it would appear to be very difficult to form such planets in stable orbits around stars of any mass if their parent nebula is very dense.

While these very preliminary results should not be overinterpreted, they are instructive because they indicate that Nakano-type disks may have difficulty forming planets in stable orbits around stars that are more than about twice the mass of the Sun. The results also indicate that planet formation from dense, compact disks is less likely to yield stable planetary systems with massive planets than is the case for formation from less dense disks. We are in the process of continuing these studies using more "realistic" models of planetary formation within a variety of disk models.

FORMATION OF THE SOLAR NEBULA

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We now have a fairly good working outline for the sequence of events that led to the formation of our solar system. Some 4.5 billion years ago, a small clump in a cloud of gas and dust moving through our galaxy began to contract under its own self-gravitation. The contraction accelerated and became a rapid collapse, with the inward motions becoming supersonic. About a million years later, the collapse had largely stopped, leaving behind the early Sun, located at the center of a rotating gaseous disk called the solar nebula. The planets of our solar system formed over the next 100 million years or so from the gas and dust of the solar nebula.

While the hypothesis that the planets formed out of the same rotating cloud as the Sun was first advanced several hundred years ago by Laplace, it has only been in the last few decades that serious work on uncovering the secrets of our solar system's formation has taken place. The Laplacian hypothesis has received strong support recently from observations of young stars similar to the early Sun, observations that have indicated the presence of rotating, gaseous disks surrounding these objects. Indeed, there really are no other hypotheses for planet formation under serious consideration at present. However, there is much that we do not yet know about how planet formation proceeded within the solar nebula.

Three approaches characterize our present efforts at understanding solar nebula evolution: (1) laboratory experiments on selected samples of the solar system, particularly meteorites; (2) astronomical observations of nearby young stars that might be analogues for the early Sun and solar nebula; and (3) theoretical modeling of the processes that might have occurred in the solar nebula. All three are crucial for establishing a self-consistent theory of the formation of our solar system; here we will describe some recent theoretical models that offer intriguing glimpses of what processes might have occurred in the solar nebula.

Theoretical models of solar nebula formation involve the solution of mathematical equations that govern the physical evolution of turbulent clouds of gas collapsing under their own weight. Because of the complexity of these equations, much of the theoretical work has been based on numerical solutions of the equations, that is, solutions that are obtained using powerful computers. Our progress in developing better numerical models of the solar nebula has largely paralleled the development of ever-faster computers. The models shown here were calculated over the last several years on the Digital VAX computers and Numerix NMX-432 array processor of the Carnegie Institution of Washington. The latter machine has recently been replaced with an Apollo DN-10040 workstation, a four processor computer that is effectively about 14 times faster on these problems. Because of the nature of the numerical code, these theoretical models are processed faster on the Apollo than on a CRAY-2 supercomputer! Fully three dimensional models of the solar nebula can easily use all this computer power and more, 24 hours a day.

The figures show several important aspects of the formation phase of the solar nebula, before the planets began to form. Figure 1 illustrates what the solar nebula probably looked like on the largest scale, as shown here about 16,000 AU across, where an AU (1 Astronomical Unit = 93 million miles) is the distance from the Earth to the Sun. The initial gas cloud has collapsed to form a high density, flattened pancake or disk, surrounded by a much lower density, fatter donut of gas that is still infalling about a million years after the collapse first started. Note the extremely low density regions (white color) directly above and below the center of the figure, where the Sun is forming. These cavities are nearly devoid of gas because with little or no rotational support, this gas falls directly onto the early Sun. Gas with substantial rotation ends up falling onto the pancake. Observations of young stars often detect oppositely directed streams of gas flowing outward from the stars in similarly shaped wind patterns; the cavities evident in figure 1 might very well have played a role in channeling our Sun's early wind. The cavities would widen with time as the wind eroded the cavity walls, possibly until

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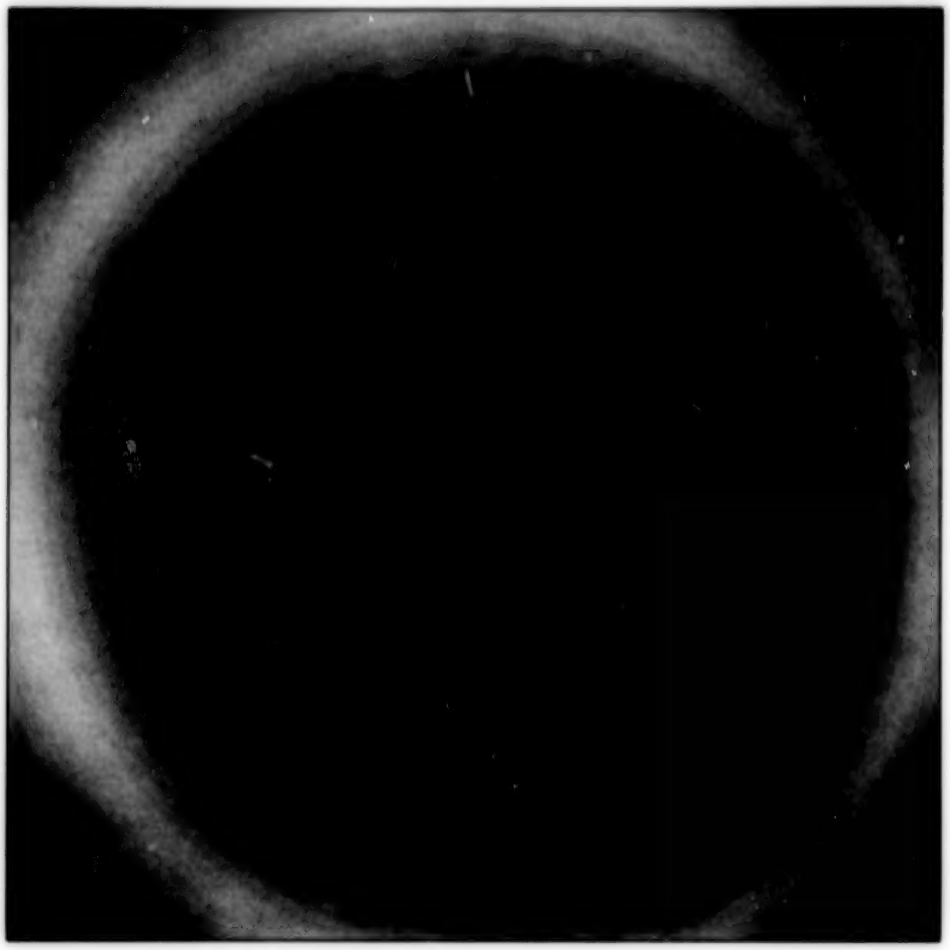

Fig. 1. Edge-on view of a computer model of the solar nebula, the cloud of gas and dust that formed our solar system. Darker colors represent higher gas densities. The nebula is rotating about a vertical axis, and has flattened into a disk perpendicular to this axis. The early Sun lies at the very center of the nebula; the early solar wind would have blown out preferentially through the "bipolar" cavities above and below the nebula.

whatever remained of the solar nebula was swept away to interstellar space.

Most of the gas in the solar nebula probably ended up in the early Sun, prior to the final clearing by the solar wind. Because of the mass of our planets, the mass of the solar nebula during planet formation must been have at least about 1% to 5% of the mass of the Sun. At even earlier times, most of the mass of the Sun may have been in the nebula, spread out to perhaps about 40 AU. One key problem in understanding the physics of the solar nebula is finding a mechanism to make most of the gas in the nebula move inward to form the Sun, while retaining sufficient gas in the nebula to form the planets, particularly the massive giant planets like Jupiter and Saturn. Several mechanisms have been suggested, such as friction (viscosity) caused by turbulent motions and magnetic fields.

Figure 2 shows a model that suggests that a third mechanism could have been important for nebula evolution. If the solar nebula became asymmetrical, for example through the formation of spiral structure similar to that in spiral galaxies, then this spiral structure can produce nebula evolution. This happens because the nebula pulls on itself through the gravitational force produced by the nebula's mass, and because the nebula is asymmetrical, gravity pulls on some regions more than others, resulting in an imbalance that moves gas inward to the central Sun, as desired.

These new computer models also predict the temperature distribution in the solar nebula, through the solution of an equation that governs the flow of radiation (principally infrared and longer wavelength light) through the nebula. This key improvement over previous models has raised the possibility of the early solar nebula

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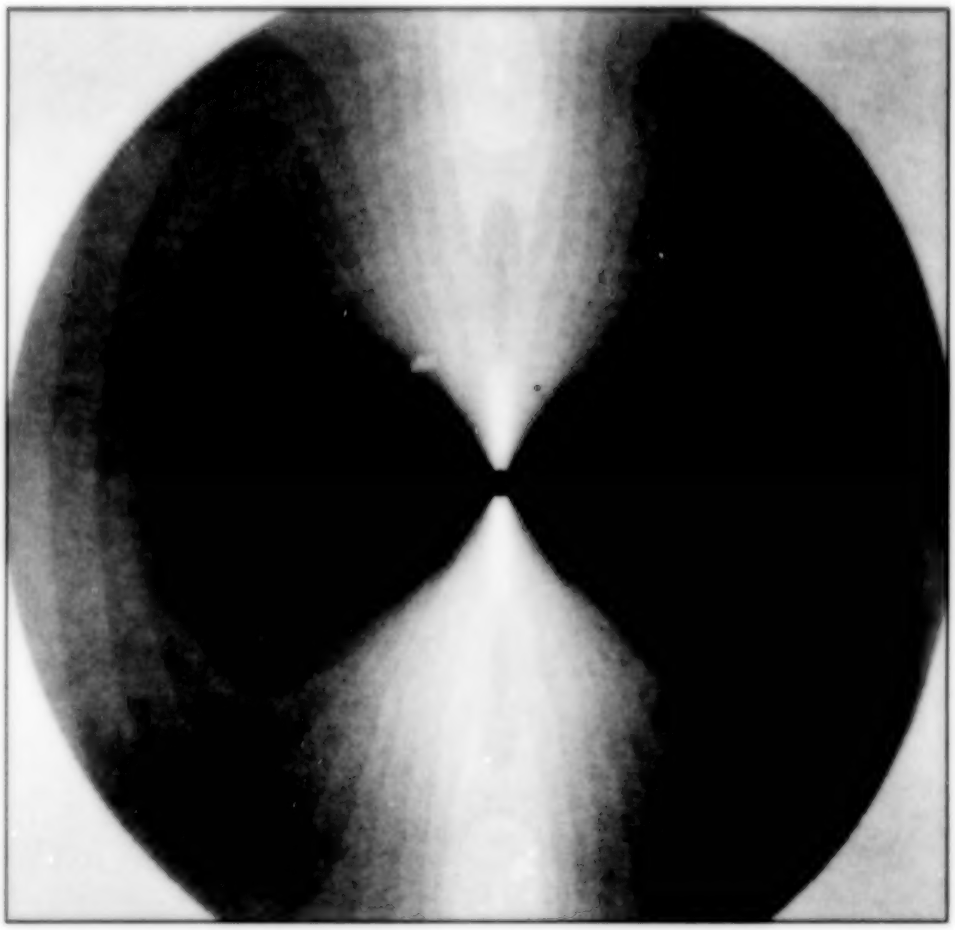

Fig. 2. Face-on view of a solar nebula model that has become significantly asymmetrical. The early Sun lies within the central bar; the bar and spiral structures evident are capable of making the nebula move onto the early Sun in about a million years. Darker colors represent higher gas densities. Region shown is 20 AU across (1 AU = Earth-Sun distance = 93 million miles).

being substantially hotter than had previously been thought to be the case. In fact, the models imply that the solar nebula may have had quite a range of temperatures, from about 2200° F at Earth's orbit, down to about −440° F at Pluto's orbit. The maximum temperature in the inner nebula (everywhere within about 4 AU of the Sun) may have been about 2200° F because of the presence of dust (iron) grains, which act like a thermostat. When solid at temperatures below 2200° F, the dust grains block the outward flow of radiation, and so the nebula heats up; above 2200° F the grains evaporate and the radiation escapes, cooling the cloud back to around 2200° F.

Such a hot inner solar nebula implies that if any of this matter was retained in the planets or asteroids, most of it must have had to cool and condense — only a few "refractory" minerals would have escaped vaporization at these temperatures. These high inner nebula temperatures imply that the inner planets would have been formed from rocky material and very little gaseous or volatile material, exactly as is observed. However, the maximum nebula temperature may have dropped precipitously to the ice condensation point (about −189° F) outside 4 AU, i.e., just inside Jupiter's orbit. This would have allowed water ice and other volatiles to remain frozen and available for planet formation in the giant planet region, and would have allowed water gas from the inner nebula to condense primarily in the area of Jupiter's orbit, which may well have aided Jupiter in becoming the largest planet. The outer planets thus may have had available abundant icy material for accumulation of their ice and rocky cores; in addition, Jupiter and Saturn evidently were able to pull in large amounts of hydrogen and helium gas from the solar nebula, prior to the final removal of the residual solar nebula.

THE EXPLORATION OF VENUS

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Venus, the morning star, has captured our attention for thousands of years. First as a celestial body to be revered, and more recently in the novels of Jules Verne and others, as a potential place for human habitation. Unfortunately, we learned from the early optical, radio and radar observations of the planet that the white clouds which enshroud the planet mask one of the most inhospitable places in the inner solar system. The clouds are mainly sulfuric acid droplets, the atmosphere is mainly carbon dioxide at a pressure ninety times that of the Earth's atmosphere and the resulting greenhouse effect has raised the surface temperature of the planet to about 1000° F. So why are we still interested in Venus? The main reason is that Venus is very Earth-like in that the two planets have very similar sizes and average densities. Studies of the atmosphere, surface and interior structure of Venus have the potential to greatly increase our understanding of the processes which formed the Earth and are controlling its evolution. Since there is no surface water on Venus, there is very little erosion and the surface of the planet has retained, for much longer than the Earth, the record of both the tectonic and impact processes that formed it.

For the past thirty years, Earth-based radar systems have been used to study the surface of Venus. By one of those fortunate coincidences, Venus is the most ideal of all the Earth-like planets for radar observations and, because of its permanent cloud cover, the one whose surface can only be extensively explored by means of radar. Over the years, improvements in both the sensitivity of Earth-based radars and in observational techniques coupled with the availability of more and more powerful computers, have allowed radar imaging of the planet's surface at higher and higher resolution. In the 1960's, the horizontal resolution was about 100 miles, in the 1970's, about 10 miles and in the 1980's, 1 to 2 miles.

In the mid-1970's, NASA provided funds to install a high powered transmitter on the 1000 ft. diameter radio telescope in Arecibo, Puerto Rico, which is operated by Cornell University for the National Science Foundation. During the close approach of Venus to the Earth in the summer of 1988, this instrument was used to obtain radar images of approximately one-quarter of the planet's surface at low northern latitudes and in the southern hemisphere. The resolution of 1 to 2 miles is approximately the same resolution as obtained by the radars on the Soviet Venera 15 and 16 spacecraft in their

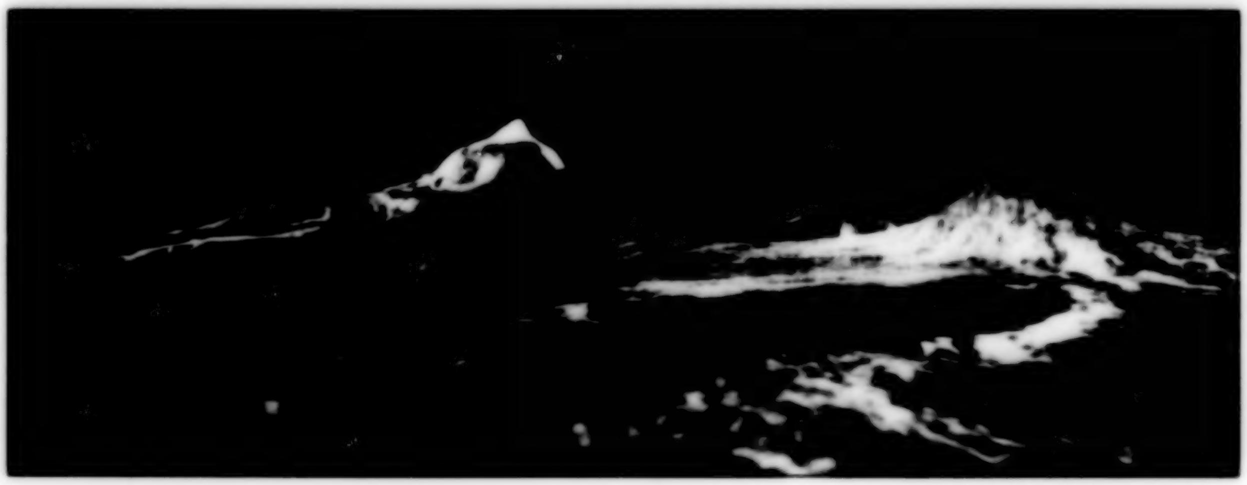

Fig. 1. An Arecibo image of two volcanoes on Venus, Sif (to the right) and Gula Montes, superimposed on the altimetry of this area obtained from the Pioneer Venus mission. The vertical extent is greatly exaggerated; Sif Mons stands about 5,000 feet above the surrounding terrain and the distance between the two mountains is 400 miles. The image shows that the bright lava flows observed in the radar image do flow downhill away from the central areas of the volcanoes (this perspective image was produced from the two data sets by B.A. Campbell of the University of Hawaii).

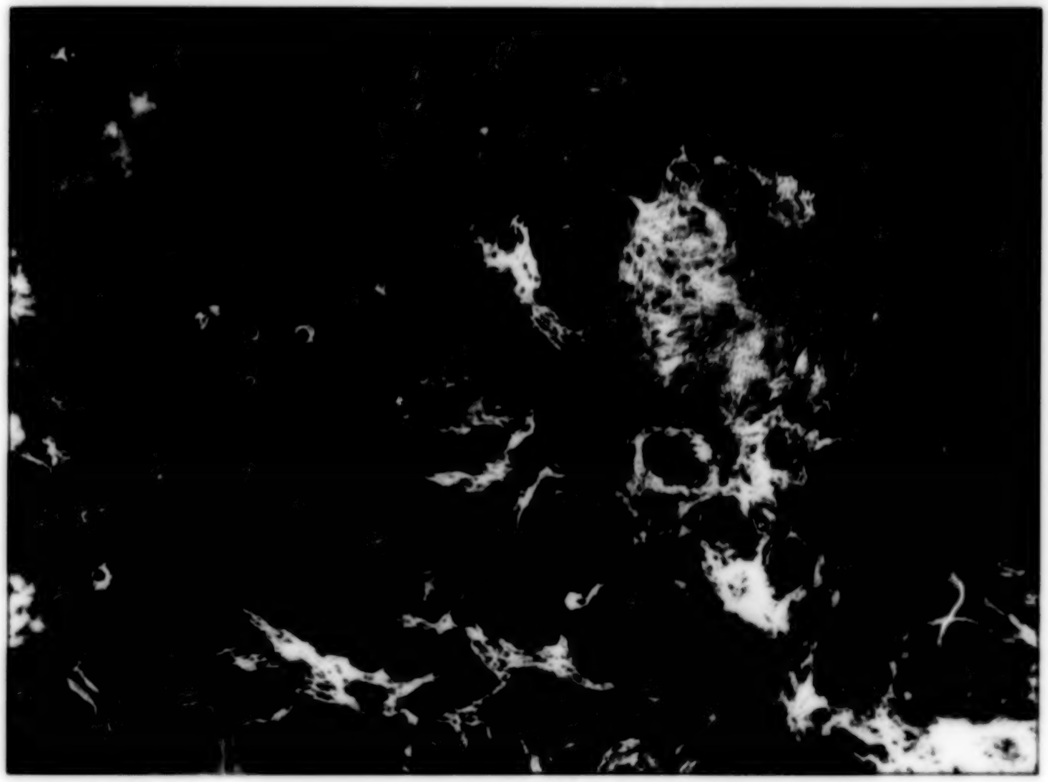

Fig. 2. A 4,000 by 2,200 mile region centered on latitude 27° south and longitude 350° east. Alpha Regio, an area of cross cutting ridges and valleys called Tessera, is at upper right and three volcanic edifices, Ushas, Innini and Hathor Montes are in a N-S line down the left hand margin. The origin of the three prominent circular features at left center is still being debated, but are probably impact craters.

survey of the northernmost quarter of the planet.

The new Arecibo data confirm the results obtained from the Soviet Venera 15 and 16 data and previous Earth-based observations. Compared with the Moon, Mercury and Mars, relatively few impact craters have been preserved on the surface indicating that it is relatively young by geological standards, with an average age somewhere between 250 and 1,000 million years. The uncertainty arises from our lack of knowledge of the rate at which impact craters have been formed on the planet over the past few hundred million years. Current evidence from the new Arecibo and Venera data indicates that the primary process responsible for the obliteration of impact craters is resurfacing due to extensive volcanic activity rather than erosional processes, as for the Earth's continents, or large scale plate tectonics, the mechanism which recycles the Earth's ocean floors. There is some still inconclusive evidence that spreading centers similar to the Earth's mid-oceanic ridges may exist on Venus but real evidence for their existence and their importance in forming the current surface of the planet will have to await results from the Magellan mission.

Several views of the surface of the planet are shown in the figures. All of the isolated mountains observed with the radar altimeter on board the U.S. Pioneer Venus Orbiter (PVO) appear in the Arecibo images to be volcanic edifices. Two of these, Sif and Gula Montes, are shown in Figure 1 where the radar image is superimposed on the altimetry data from PVO. Figure 2 shows an area approximately 4,000 by 2,200 miles just south of the equator. Alpha Regio, the large high contrast feature at the upper right, is an area where the surface has been deformed into a series of cross cutting ridges and valleys by a process which we do not yet understand. Three volcanic edifices lie along the left hand margin of the image, again corresponding to mountains in the PVO data. A number of circular crater-like features can be seen, some of which are thought to be impact craters while others may be of volcanic origin.

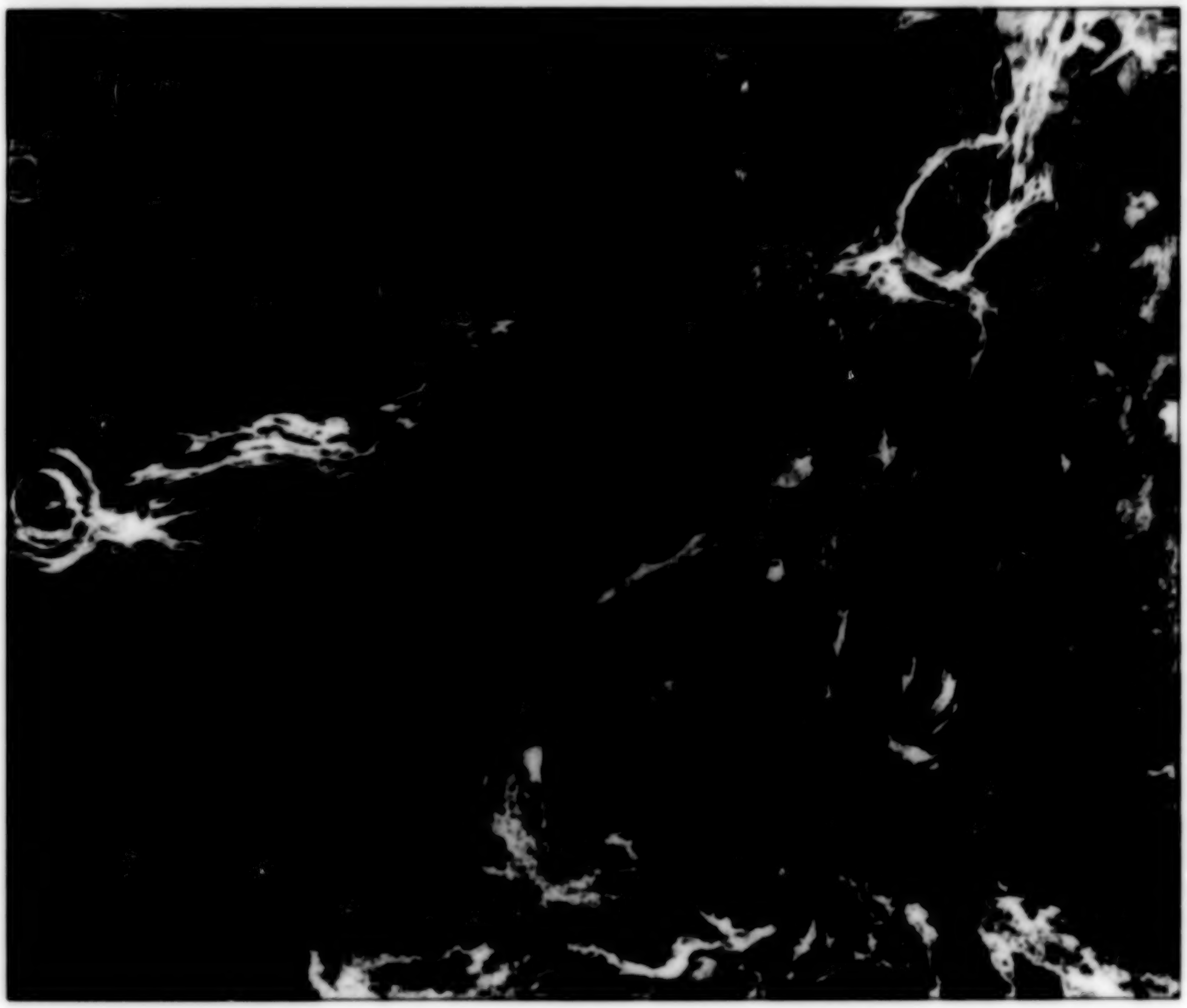

Fig. 3. This Arecibo radar image has provided our first look at this region at high southern latitudes on Venus. It covers an area 2,000 by 1,500 miles centered on latitude 62° south and longitude 350° east. The double-ringed structure at upper left is thought to be an 80 mile diameter impact crater, which has been named after the great woman physicist, Lise Meitner. The other three circular structures are thought to be the result of convective processes in the interior of the planet. An extensive volcanic field can be seen at upper center.

A number of large ovoidal structures normally 200 to 400 miles across were discovered in the Soviet Venera data and three examples can be seen in the Arecibo image in Figure 3, which covers an area 2,000 by 1,500 miles at high southern latitudes. These features, which have been named Coronae, are thought to be the surface expression of convective processes in the interior of the planet. The Corona in the lower right of the image, at about 500 miles across, one of the largest so far discovered, shows evidence for significant volcanic activity in its center. The smaller double-ringed structure in the upper left hand corner of the image is thought to be a large double-ringed impact crater about 80 miles in diameter.

The Magellan mission to Venus will provide radar images of the planet's surface at resolutions of 500 to 850 feet, 10 times better than the results from the Arecibo and Venera observations. These latter observations, plus the altimetry data from the Pioneer Venus mission, provided us with an initial view of the planet's surface, and a base from which the results the Magellan mission can be understood.

CREEP OF PLANETARY ICES

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While the inner planets and their satellites are mostly rocky, the outer planets and their satellites, beginning at the orbit of Jupiter, are made up of rock plus a certain amount of condensed (i.e., liquid and solid) volatiles such as water, ammonia, and methane. On many of the satellites of Jupiter, Saturn, Uranus, and Neptune, the volatiles are plentiful enough to profoundly influence the evolution of the satellites. For instance, Jupiter's giant moons Ganymede and Callisto are well over half water by volume, probably most of it frozen. The dramatic images returned by the Voyager spacecraft of the surfaces of these two satellites show the effects of water ice creeping, flowing, and breaking under the influence of tectonic forces like those that build mountains and cause continents to drift on our own planet. The Voyager photos show that active tectonics occurs (or did occur) on icy satellites throughout the outer solar system. Even at its coldest reaches, on Neptune's enigmatic giant moon Triton, signs of a stretched, crept, and otherwise deformed surface are ubiquitous.

The mechanical properties of ices play a key role in determining how internal forces affect the evolution of icy moons. The spacing and depth of surface faults depends on how easily ices fracture under forces that push and pull. The strength of ice under ductile conditions (i.e., in the absence of fracturing; so-called "plastic" flow) determines the rate of topographic relaxation (of, say, craters) and influences the likelihood of internal convection, which in turn strongly influences the possibility of planetary-scale differentiation. Differentiation is usually the result of a thermal runaway, which occurs when a gravitationally unstable mixture (like ice and rock) starts to melt. Paradoxically, the stronger the ice, the greater the possibility of planetary-scale melting and differentiation. The reason for this is that the main source of heat is usually internal to a planet: the radioactive decay of certain elements in the rocks. If heat can be dissipated to the surface, say by solid-state convection of ice, melting will not occur. However, if ice is resistant to deformation, dissipation of internal heat will not be efficient, internal temperatures will rise, and melting and differentiation will occur.

The mechanical properties of ices can be determined by experiment, and we are conducting such experiments in joint research at our laboratories. The apparatus shown in Figure 1 is a high-pressure, low-temperature deformation rig designed to test the strength of frozen and partially melted volatiles at temperatures and pressures found on and within the icy satellites. Hydrostatic pressures within the steel vessel can reach 600 MPa (about 90,000 psi) (14.7 psi = atmospheric pressure), simulating pressures to deep within the mantles of the icy giants Ganymede, Callisto, and Triton, and throughout the interiors of the smaller icy moons of Saturn and Uranus. Environmental temperatures are controlled to as low as 77 K (-321° F) by immersing the pressure vessel in a cold bath of liquid nitrogen or liquid or frozen ethanol, thus simulating the coldest surface temperatures on all the icy moons with the possible exception of Triton, which may be a few degrees cooler.

The only environmental variable we can not simulate is the imposed deformation (strain) rate, a problem inherent in all geological tests of this sort. Large deformations are possible in geologic time at exceedingly low deformation rates. In the laboratory, we have only days or weeks to produce the same deformations, so we must push the samples unnaturally hard to make them deform unnaturally fast. For instance, we can shorten our 2.5-inch-long ice cylinders as slowly as 0.050" (2%) per day. This is still at least 5 orders of magnitude faster than solid-state convection runs in the interiors of planets. We hope on the basis of understanding the physics of deformation that it is valid to extrapolate our results to the slower deformation rates.

To date, we have tested pure water ice, ices of ammonia plus water, methane plus water, and mixtures of water ice plus silicate dust. In addition to the strength of the "ordinary" ice phase called ice I, we have measured the strength of several of the exotic high-pressure, high-

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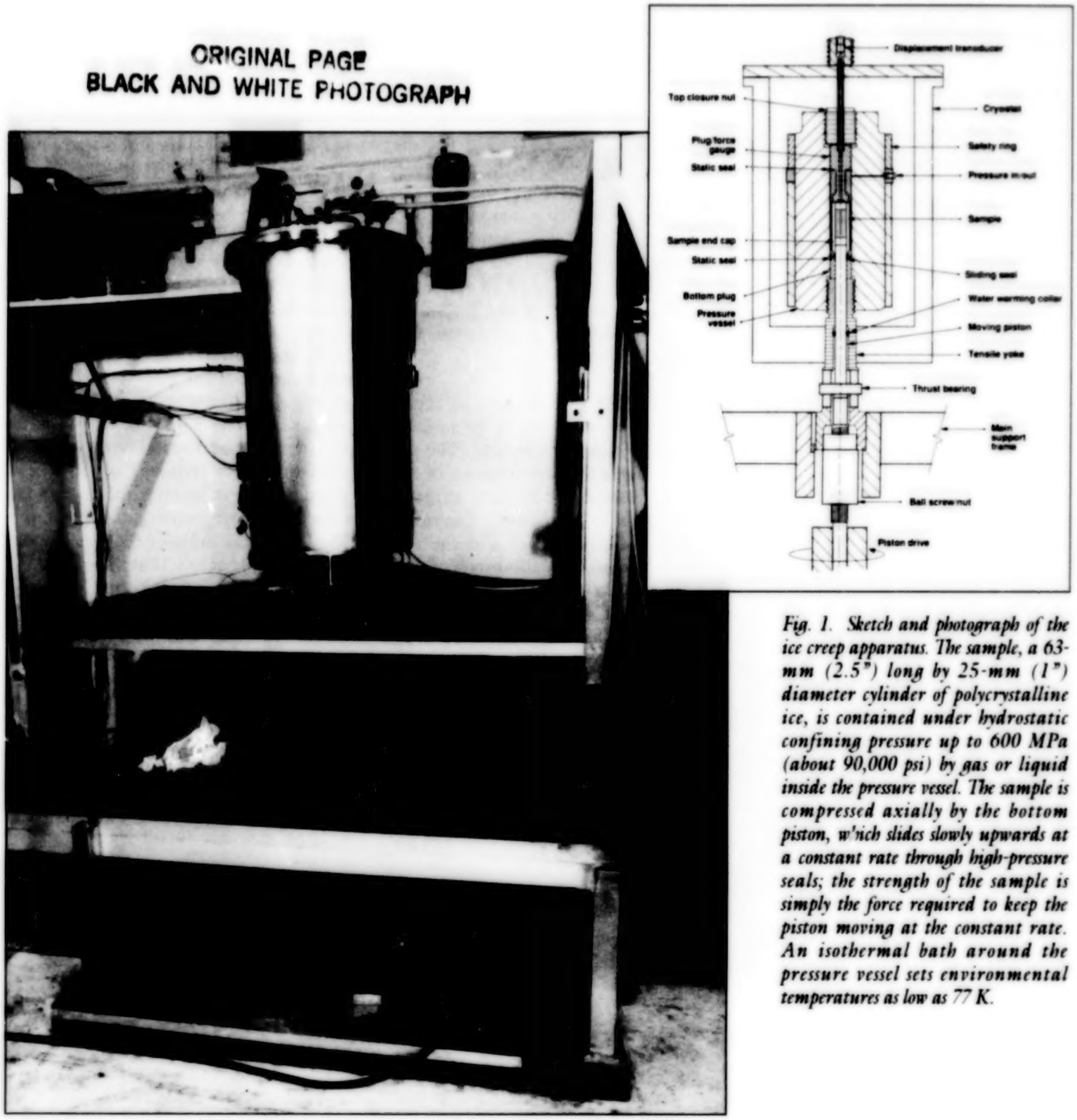
Displacement transducer
Top closure nut
Cryostat
Plug/force gauge
Safety ring
Static seal
Pressure in/out
Sample
Sample end cap
Static seal
Sliding seal
Bottom plug
Water warming collar
Pressure vessel
Moving piston
Tensile yoke
Thrust bearing
Main support frame
Ball screw/nut
Piston drive

Fig. 1. Sketch and photograph of the ice creep apparatus. The sample, a 63-mm (2.5") long by 25-mm (1") diameter cylinder of polycrystalline ice, is contained under hydrostatic confining pressure up to 600 MPa (about 90,000 psi) by gas or liquid inside the pressure vessel. The sample is compressed axially by the bottom piston, which slides slowly upwards at a constant rate through high-pressure seals; the strength of the sample is simply the force required to keep the piston moving at the constant rate. An isothermal bath around the pressure vessel sets environmental temperatures as low as 77 K.

density phases of water ice (namely, ices II, III, and V) that are expected to exist deep in the interiors of the icy giants. Some of the results for ductile flow are compared in Figure 2. Figure 2 shows that all ices get stronger with decreasing temperature, that ice + silicate dust is stronger than ordinary ice, that the presence of ammonia weakens water ice considerably (especially as one approaches 176 K (–143° F), where a partial melting occurs), that $CH_4 \cdot H_2O$ (called methane clathrate) has about the same strength as water ice, and that ice II is slightly stronger than ice I. Not shown in Figure 2 is ice V, which is slightly weaker than ice I, and ice III, which is a great deal weaker.

Our data are applied quantitatively to the icy moons through the use of computer models, although we can see some qualitative indications directly. For instance, the low strength of ammonia-water ices helps explain tectonic activity on the small moons of Saturn and Uranus, which are expected to be relatively ammonia-rich, but which have cooler interiors than the icy giants. Analysis of surface topography suggests that the outermost layers of Ganymede and Callisto are probably dirty ice: based on our data, a surface layer of pure ice would result in crater walls and basins relaxing more than they actually have.

Our results leave us tantalizingly close to an answer to the so-called Ganymede/Callisto dichotomy. The two moons are almost twins in terms of size and composition, but their external appearance is very different. Ganymede has more young surface and signs of extensional tectonics (rifts, grabens, and the like) than Callisto. It has been suggested that Ganymede differentiated while Callisto did not, the result perhaps of a subtle instability during the evolution of the planet. Ice does deform readily at planetary temperatures and pressures, we now know, but perhaps not readily enough to suit Ganymede's heat engine. Perhaps the unusual weakness of ice III (at a depth of a few hundred kilometers) played a role.

We have also learned a great deal about the behavior of ice in the brittle field, where deformation occurs by fracturing or by sliding on preexisting fractures. Ice samples break at lower loads than silicates, but the depth of fracturing (sometimes called the depth of the brittle-to-ductile transition) on, say, Ganymede is still limited to about 3-6 miles by the low plastic flow strength of ice relative to silicates. In other words, below 3-6 miles the stress required to break ice is higher than the stress required to deform it plastically. It is interesting that under increased hydrostatic confinement the strength of ice samples that are already faulted approaches that of unfaulted samples, so the 3-6 mile brittle-ductile transition depth may not be strongly influenced by whether or not a fracture already exists. The presence of silicate dust in the ice does not increase the fracture strength (as it does the the plastic flow strength) but it does promote ductility. The effect of dirt in the crust of Ganymede would thus be to raise the brittle-ductile

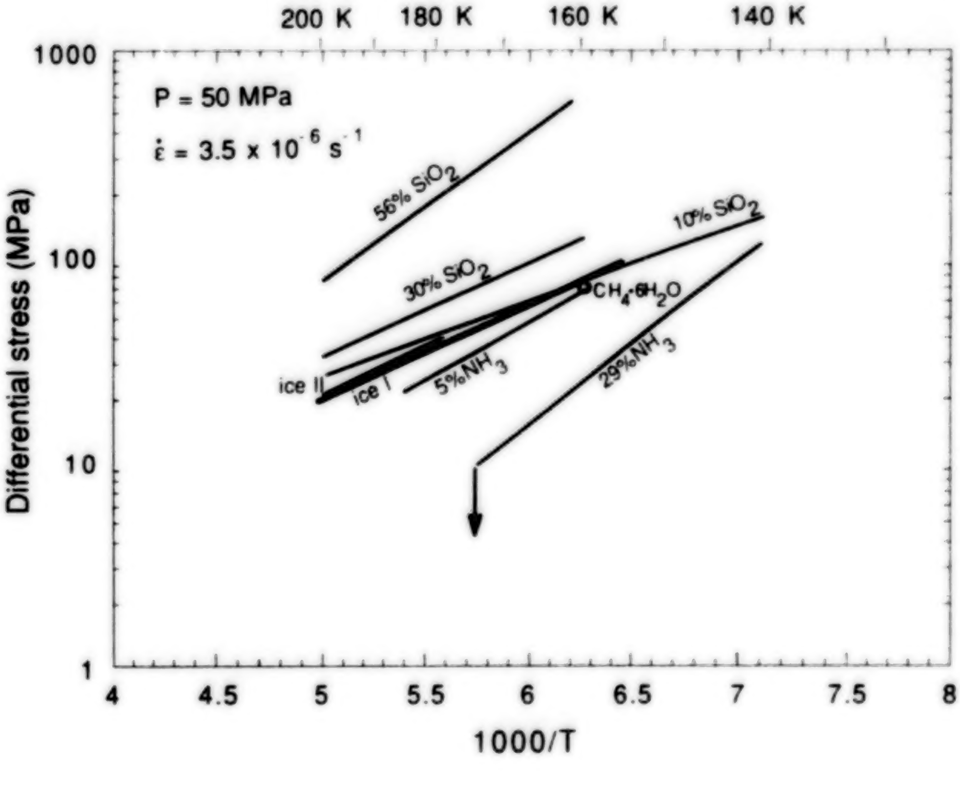

Fig. 2. Strength of ice (i.e., differential stress supported by a sample) vs. temperature and inverse temperature for various types of ices. Results are shown for pure water ices I and II, for water ice with three different volume percentages SiO_2 sand, for water plus two different weight percentages ammonia (NH_3), and for methane clathrate ($CH_4 \cdot 6H_2O$).

HOW OLD ARE THE UREILITE METEORITES?

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Igneous meteorites, called achondrites, are rocks that in many ways resemble igneous rocks on the Earth. (Igneous rocks are rocks that crystallized from silicate liquids.) They contain many of the minerals found in terrestrial igneous rocks, and these minerals have forms and spatial relations that are similar to those of their terrestrial counterparts. Thus, igneous meteorites appear to be products of igneous differentiation processes (that is, melting, crystallization, and physical separation of crystals and liquid) on other planetary bodies.

Determining the ages of igneous meteorites is extremely important, because when an igneous meteorite formed can indicate something about the size, identity, and history of its parent body. In order to undergo igneous differentiation, a planetary body must have an internal source of heat sufficient to cause rocks to melt. This heat is supplied primarily by the residual energy of accretion of the planets 4.55 billion years ago and by decay of naturally occurring radioactive elements. The planets are continually losing heat, however. In general, the smaller the planetary body, the more quickly heat was lost after accretion, and the shorter the duration of igneous processes. On the Moon, which has a radius only one-quarter that of the Earth, igneous processes had virtually ceased by about 2.5 billion years ago. On the Earth, lava flows and explosive volcanic eruptions like that of Mount St. Helens are ample evidence that igneous processes are still occurring today.

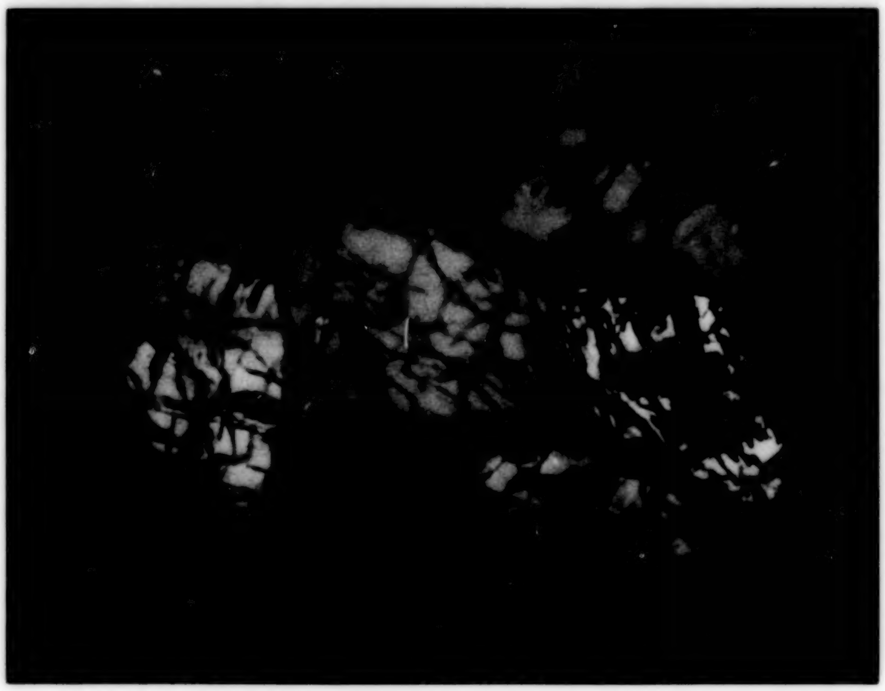

Fig. 1. A view of a thin slice of Antarctic ureilite ALHA77257, as seen through a microscope with crossed polarizers. The major minerals are olivine (iron-magnesium silicate) and pyroxene (calcium-iron-magnesium silicate).

The most abundant igneous meteorites are the basaltic achondrites, or eucrites. These have been found to be very old. The youngest age measured on a eucrite is about 4.4 billion years. The very old ages of these rocks suggest that they originated on a body even smaller than the Moon. One widely held hypothesis is that the eucrites come from Vesta, which is the third largest asteroid.

Another very interesting group of igneous meteorites is referred to as the SNC group (shergottites-nakhlites-chassignites). These rocks have ages at least as young as 1.3 billion years, and some workers believe they formed as recently as 180 million years ago. These relatively young ages indicate that the SNC meteorites formed on a relatively large parent body. This finding is one of the most important pieces of evidence supporting the popular hypothesis that the SNC meteorites come from Mars.

The second most abundant igneous meteorites are the ureilites. Ureilites are ultramafic rocks (rocks very rich in the elements iron and magnesium). They consist mostly of the minerals olivine (iron-magnesium silicate) and pyroxene (calcium-iron-magnesium silicate), and they resemble rocks from the Earth's mantle (Fig. 1). Studies of the bulk compositions of these rocks and the compositions, forms, and spatial relations of their minerals have shown that they formed by one of two types of igneous processes. The first alternative is that they are solid residues left behind after extensive, incomplete melting and physical extraction of the melt. The second

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alternative is that they are aggregates of crystals that formed in a silicate melt and were physically concentrated. In either case, the evidence suggests a parent body that has experienced extensive igneous activity and may therefore be fairly large.

A serious contradiction in the data arises, however, when the oxygen isotopic characteristics of the ureilites are also taken into account. The oxygen isotope patterns of ureilites, recently measured by Robert Clayton and his co-workers (University of Chicago), indicate that these rocks are not related to one another by igneous processes. Rather they seem to be primitive material which has not been fractionated relative to bulk solar-system material. This contradiction, between one set of evidence indicating extensive igneous evolution and the other set indicating no igneous evolution, is not encountered for any other group of known meteorites. Determining the cause of this contradiction constitutes the biggest obstacle to understanding the geologic history of the ureilite parent body.

Until recently there were no data on the age (or ages) of ureilites. Age information could be critical to resolving the question of whether ureilites are primitive material or are the products of complex and extended planetary igneous processes. With this in mind, my co-workers P. Jonathan Patchett, Michael J. Drake (University of Arizona, Tucson) and I undertook to determine the ages of these rocks, using the samarium-neodymium isotopic method. The project was a potentially very difficult one, because the concentrations of neodymium and samarium in ureilites are extremely low, hence the isotopic compositions of these elements are very difficult to measure precisely. We first analyzed whole-rock samples of several different ureilites. Whole-rock analyses give first-order age information. Analyses of concentrates of the individual minerals (mineral separates) gives more information but is technically much more difficult.

The results of our whole-rock analyses showed that seven samples taken from three different ureilites (Kenna, Novo Urei, and the Antarctic meteorite ALHA77257) plot on a line on the standard samarium-neodymium evolution diagram (Fig. 2). When whole-rock sample data define a linear array on this type of diagram, such an array can have two possible interpretations. The first possibility is that the samples are all related and all formed at the same time. In this case the linear array is called an isochron and its slope gives the time the rocks formed. If the array on Figure 2 is an isochron, the rocks formed 3.74 billion years ago. The second possibility is that the samples consist of mixes of varied proportions of two unrelated end member components. In this case, the linear array is called a mixing line, and its slope has no age significance.

For our data, we originally favored the interpretation that the array is a mixing line, for two reasons. First, mineralogic studies indicate that ureilites do indeed appear to be mixtures of two components. One is the olivine plus pyroxene assemblage that makes up the bulk of the rocks, and the other is a minor, unidentified component with an odd composition. The former component has been shown to have a very high ratio of samarium to neodymium and the latter has been shown to have a very low ratio of these elements, so these components could easily represent the two end members of a samarium-neodymium mixing line.

The second reason we favored a mixing interpretation is that we found that whole-rock samples of three other ureilites (Antarctic meteorites PCA82506, ALHA82130, and META78008), which appeared to be devoid of the second component, had ages of 4.55 billion years (as determined by the samarium-neodymium method). This finding suggested to us that the olivine plus pyroxene assemblage of the ureilites formed 4.55 billion years ago, and that at some later, undetermined time, the second (low samarium/neodymium) component was introduced into the rocks. Our interpretation supported the hypothesis that ureilites are relatively primitive materials rather than being the product of an extended history of igneous processes.

The next step of our project was to test our hypothesis by determining whether the linear array shown in Figure 2 is a mixing line or an isochron. To do this, it was necessary to produce quite pure concentrates of the individual minerals in the ureilites and analyze these. We chose to obtain a pyroxene concentrate because pyroxene has the highest neodymium content of any of the major minerals in ureilites. If the linear array is a mixing line, as we had hypothesized, the pyroxene should have an age of 4.55 billion years. Alternatively, if the linear array is an isochron, the pyroxene point should plot on that line, indicating that it, as well as all the other samples that plot on the same line, came to chemical equilibrium 3.74 billion years ago.

The task of creating a pure pyroxene concentrate was a long, tedious one. Pyroxene grains could only be reliably separated from olivine and the other minor constituents of the rocks by visual identification, so we ground a sample of the ureilite Kenna to a very fine powder (70-100 micrometers) and picked out the fragments of pyroxene grains under a binocular microscope. It took about ten months of hand picking before we had minimally enough pure pyroxene to analyze. Because of the very small mass of the concentrate, we analyzed it at the University of California, La Jolla, where Dr. Gunter Lugmair has developed techniques for precise analysis of very small amounts of neodymium.

The result of this analysis (Fig. 2) showed that, in contrast to our expectation, the pyroxene plots on the 3.74 billion year line. Thus, this line must be an isochron that gives the age of the three ureilites Kenna, Novo Urei, and ALHA77257. The geologic interpretation of a 3.74 billion year age for these rocks is, however, still not clear cut. The simplest possible interpretation is that they crystallized from a silicate melt at this time, and this possibility cannot be ruled out. However, this interpretation is not consistent with our observation that the three other ureilites we studied have ages of 4.55 billion years. Another possible interpretation is that Kenna, Novo Urei, and ALHA77257 crystallized 4.55 billion years ago, but they were wholly or partly remelted 3.74 billion years ago. When they were remelted, they came to chemical equilibrium with a material having a low ratio of samarium to neodymium. Distinguishing between these two possibilities may depend upon identifying and understanding the nature of the low samarium/neodymium component of ureilites, a task which constitutes our next goal.

Thus, our study has so far not produced a simple answer to the question "How old are ureilites?", nor has it resolved the apparent contradiction between the oxygen isotopic characteristics of these rocks, which indicate they are primitive, and the mineralogic characteristics, which indicate they are not primitive. However, our work has shown that ureilites cannot be simply primitive materials that formed 4.55 billion years ago and remained undisturbed since that time. Rather, they have had a complex chemical and physical history, and they must have formed on a parent body that experienced igneous activity at least as recently as 3.74 billion years ago.

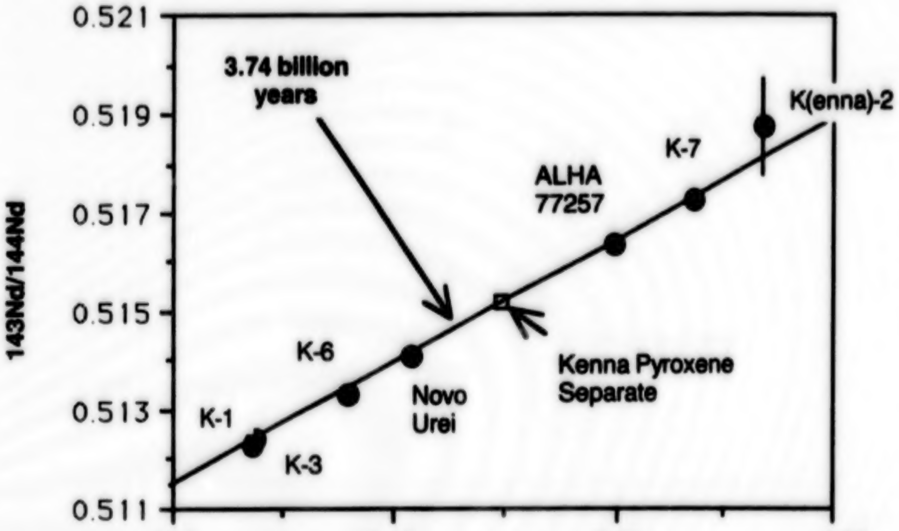

Fig. 2. Standard samarium (Sm) - neodymium (Nd) isotopic evolution diagram, showing linear array formed by analyses of whole-rock samples of three ureilites — Kenna (K), Novo Urei, and Antarctic ureilite ALHA77257. A pyroxene concentrate (separate) plots along this linear array, indicating that the line is an isochron (see text) and that the age of the three ureilites is 3.74 billion years.

EVIDENCE FOR MAGMATIC WATER ON MARS

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One of the many enigmatic features on the surface of Mars is the system of valleys that seems to be the result of water erosion on a planet now containing very little water. It is widely believed that Mars once contained much more water than it now does, but independent evidence has been understandably lacking. The discovery within the last ten years that certain meteorites may in fact be pieces of Mars, ejected from its surface by giant impacts long ago, promises to provide tangible clues to the martian water budget in the distant past.

These particular meteorites are igneous rocks, formed by solidification of molten rock in a subterranean environment where they cooled slowly. While other meteorites thought to have formed in this manner exist, what distinguishes the "martian" ones are their relatively young ages: virtually all other meteorites are 4 1/2 billion years old, as old as the solar system itself, but a select small group (eight are now known) are much younger (less than 1 1/3 billion years). For a planet to have such recent igneous activity requires either that it be large enough (like Earth) to prevent rapid cooling by conduction, or else have an ongoing external energy source (e.g. Jupiter's moon Io). Earth's Moon, for example, long ago cooled to the point where igneous activity effectively ceased. A martian origin for the eight "young" meteorites is suggested by careful analysis of rare gases, trapped within the rocks when the giant impact that sent them flying into space caused small pockets of the rock to melt and dissolve atmospheric gas from the parent planet. A comparison of these gases with an analysis of the martian atmosphere by the Viking spacecraft shows striking similarities.

Igneous rocks can provide much insight about the interior of the planet, but independent knowledge is required about the chemical compositions both of the minerals in the rocks and the melt from which they formed. The latter is the more difficult to ascertain. On Earth, it is often possible to find out directly, or at least make a reasonable estimate, of the melt composition. Unfortunately, this is not so easy for the meteorites in question because these rocks formed deep within the parent planet, where crystals sank under the influence of gravity and became separated from the liquid they crystallized from. The solution to the problem has been to recognize and analyze small pockets of melt that became trapped within the growing crystals (Fig. 1). These melt inclusions can either quench to a glass or else crystallize, but because they are isolated from what is going on outside of the crystal, they provide a frozen-in sample of melt at an early stage of its evolution. Detailed examination of melt inclusions in several of the martian meteorites has given information about the martian interior at the time the rocks formed. We now know, for example, that the martian melt compositions were broadly similar to a common type of volcanic rock found on Earth (basalt) but richer in iron and poorer in aluminum.

A more exciting discovery is that the martian magmas contained significant quantities of dissolved water. This conclusion is based on the observation that the melt inclusions contain minerals, such as mica and amphibole, in which water is an essential component. In fact, laboratory experiments have shown that the melts from which those phases formed must have had at least 4-5 wt.% dissolved water. Therefore the trapped melt inclusion must have had roughly this same water content. This is strong evidence that Mars contained magmatic water as recently as a billion years ago.

This conclusion has important implications for the evolution of the martian atmosphere. For example, the fact that so much water was present in the martian interior 3 1/2 billion years after the planet formed means that the martian volatile inventory (e.g. gases, water) is more complicated than previously suspected. If all of the volatiles that were originally contained in the planet were outgassed immediately after its accretion 4 1/2 billion years ago, then some process has acted to recycle volatiles back into the martian interior. Alternatively, perhaps the planet did not outgas completely and some volatiles were retained in the interior, being slowly leaked to the surface by volcanic eruptions. The very nature of that volcanism is itself strongly dependent on the presence of water in the melt. The physical properties of erupted magma (e.g. viscosity), its chemical composition, and the style of volcanism (explosive or quiescent) all depend on the water content of the melt at depth and on the surface. And, of course, the ability of a planet to support life is critically dependent on the availability of liquid water. The results of these studies suggest that liquid water did exist on Mars in the past, thus giving credence to the possibility that Mars may once have supported life.

Fig. 1. Photomicrograph of a melt inclusion in one of the martian meteorites. Dark gray material is glass (melt). Scale bar is 100 microns (~0.04 in).

THE ANGRITE METEORITES: ENIGMATIC ANCIENT MELTS OF SOLAR SYSTEM MATERIAL

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The angrite meteorites are rare and enigmatic. These stones are "basaltic" meteorites (basalts are solidified lavas rich in the elements iron and magnesium). They have unusual elemental compositions and their primitive isotopic compositions indicate they are very ancient solar system material. Only three angrites have been found (Fig. 1). The first fell in Brazil in 1869 and is referred to as Angra dos Reis (abbreviated AdoR). The other two were found in Antarctica in 1986 and 1987.

All three angrites contain the minerals pyroxene (calcium-magnesium-iron silicate) and olivine (magnesium-iron silicate), and the Antarctic angrites also contain the mineral anorthite (calcium-aluminum silicate). These minerals are of similar, odd compositions in all three stones. The pyroxenes, in particular, are very unusual. They are much richer in aluminum and titanium than pyroxenes in most other types of rocks. The olivines also are unusual in that they are high in calcium. No other "basaltic" meteorites have similar pyroxene as a major component or have calcium-rich olivine.

What is the origin of these strange rocks and how are they related to each other? Are angrites truly rare in the solar system or is their rarity on Earth simply an accident of sampling? Are the primitive isotopic compositions related somehow to the strangeness of the angrite elemental compositions?

Early results. The first concerted effort to understand what at that time was the only angrite was made in the mid 1970's by Klaus Keil (then at University of New Mexico, Albuquerque; now at University of Hawaii, Honolulu), who organized a consortium effort to decipher the history of Angra dos Reis. The work of this group led to two major conclusions. First, Marty Prinz (then at University of New Mexico, now at American Museum of Natural History, New York) and his co-workers concluded that AdoR, which consists almost entirely of pyroxene, formed by physical concentration of pyroxene crystals that grew from a silicate melt. Second, Gerald Wasserburg and his co-workers (California Institute of Technology, Pasadena) concluded that AdoR is very old indeed — the pyroxene in this rock crystallized about 4.6 billion years ago, at about the time the solar system formed. This conclusion is based on measurement of ratios of the isotopes of strontium. Wasserburg and his co-workers found that the ratio of strontium-87 to strontium-86 in the melt when the pyroxene crystals formed was the lowest such ratio ever measured in a basaltic rock. Because strontium-87 is produced by the radioactive decay of rubidium-87, the amount of this isotope is continually increasing, whereas the amount of strontium-86 remains constant. Therefore, the older the rock, the less strontium-87 there was when it formed, and the lower its original ratio of strontium-87 to strontium-86.

Rethinking the paradigm: the next ten years. Questioning previously accepted viewpoints is one of the cornerstones of the scientific method. Therefore, it was natural that there should be some disagreement with the model for the formation of AdoR outlined above. In 1982, John Jones (then at California Institute of Technology) noted that AdoR is comparatively rich in uranium, even though the element uranium is not incorporated in the mineral pyroxene in significant amounts. Jones suggested that some of the pyroxene in the rock actually represents solidified silicate melt, because uranium tends to concentrate in silicate melts. In 1989, Allan Treiman (Boston University) took this notion one step farther and suggested that a very large proportion of the pyroxene in the rock represents solidified melt. He proposed that, in fact, the composition of the melt was very close to the composition of the pyroxene. A variety of experiments by Treiman, Gary Lofgren (NASA Johnson Space Center) and Edward Stolper (California Institute of Technology) indicated that this hypothesis was indeed possible.

In 1988, another model for the origin of AdoR was presented by Marty Prinz. Prinz noted that the odd composition of the pyroxene in AdoR resembles the composition of pyroxene in a type of rock

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found as fragments included within some carbonaceous chondrites (chondrites are primitive meteorites unaffected by melting processes that occurred in planetary or asteroidal bodies). The rock in these fragments has a very high melting temperature and is very rich in calcium and aluminum; the fragments are commonly referred to as refractory inclusions. Prinz hypothesized that a parent rock rich in refractory inclusions was melted to form the silicate liquid from which AdoR subsequently crystallized. Prinz viewed AdoR as primitive matter that was only one or two steps removed from the beginning of the solar system.

Thus, by the end of the 1980's, several models for the origin of AdoR had been proposed. Although no consensus had been reached, it was generally recognized that AdoR's past was complex.

More angrites: the Antarctic connection. The two angrites found in Antarctica are grouped with AdoR, not because they are identical to it (far from it; see Fig. 1), but because the chemical compositions of the minerals in these rocks is like that in AdoR. Instead of being composed almost entirely of pyroxene like AdoR, however, the first Antarctic angrite found (Lewis Cliff 86010) consists of approximately equal amounts of pyroxene, olivine, and anorthite. The second Antarctic angrite (Lewis Cliff 87051) is richer in olivine than the other angrites. This stone consists of many crystals of olivine within a matrix of much smaller crystals, and this matrix has a bulk composition similar to that of Lewis Cliff 86010.

It is interesting to note that, because of our past experience analyzing other meteorites and lunar samples, we can now extract a lot of information from very small samples. This is important in the case of Lewis Cliff 87051 because this sample weighs only about 0.6 gram (1/50 of an ounce; see Fig. 2). With only a half gram of rock, it is possible to: obtain a bulk chemical composition for major, minor and trace elements; provide a detailed description of the compositions, forms and spatial relationships of the minerals; determine the

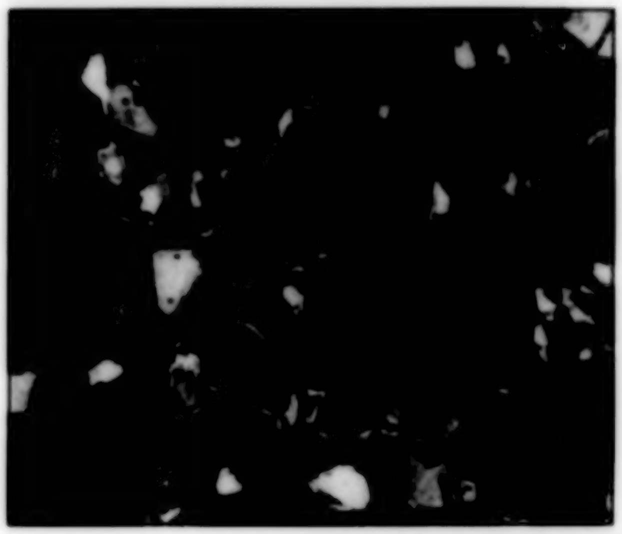
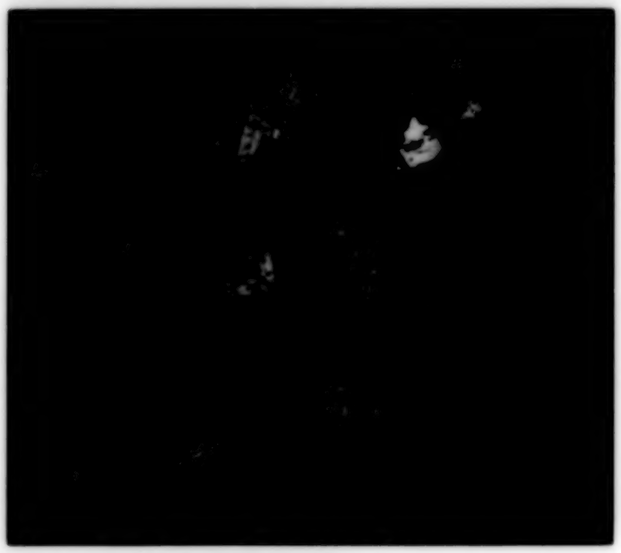


Fig. 1. Photomicrographs of the three angrites. (a) Transmitted polarized light photo of Angra dos Reis. The large brown crystals are thought to be part of a single crystal (continuous in 3-D) that grew in a dendritic (tree-like) manner (photo courtesy of A.H. Treiman).



(b) Transmitted polarized light photo of LEW86010 (Big LEW). Pyroxene, olivine and anorthite are all present in this photo. Note the difference in texture between this photo and (a) and (c).

(c) Backscattered electron photo of LEW87051 (Little LEW). Again the texture is different from either AdoR or Big LEW. Large crystals are calcium-rich olivines.

age of formation; and still have material left for future generations of scientists to study!

New insights from the new angrites. What have we learned from these new angrites? First, several investigators, including Larry Nyquist (NASA Johnson Space Center) and Gunter Lugmair (University of California, San Diego) and their co-workers, have discovered that the initial isotopic composition of strontium in the angrites, although primitive, is not quite as primitive as was once thought. These new analyses indicate that the strontium isotopic composition of the angrites is indistinguishable from that of the eucrites, another group of basaltic meteorites. Second, Robert Clayton (University of Chicago) has determined that the oxygen isotopic composition of Lewis Cliff 86010 agrees with that of AdoR. This observation supports a relationship between the two rocks, because oxygen isotopic ratios should be similar in meteorites that are related. Interestingly, the oxygen isotopic composition of the angrites is indistinguishable from that of the eucrites. Third, Gordon McKay (NASA Johnson Space Center) and Ghislaine Crozaz (Washington University, St. Louis) and their co-workers have found, by using data on the chemistry of europium, that Lewis Cliff 86010 formed under relatively oxidizing conditions. Europium is an element that can change its chemical behavior, depending upon how oxidizing the ambient conditions are. McKay and Crozaz and their co-workers compared the results of laboratory experiments with ion microprobe analyses of the minerals in Lewis Cliff 86010 and have concluded that, when this rock crystallized, conditions were about 100 times more oxidizing than when the eucrites crystallized.

More insights from experiments. Recently, we have performed experiments at Johnson Space Center that help resolve the problem of the origin of the angrite melts. We and Amy Jurewicz have performed partial melting experiments on a carbonaceous chondrite, the Allende meteorite (about a ton of this meteorite fell at the little town of Pueblito de Allende in Mexico, so we don't feel guilty about melting a few grams). The experiment was

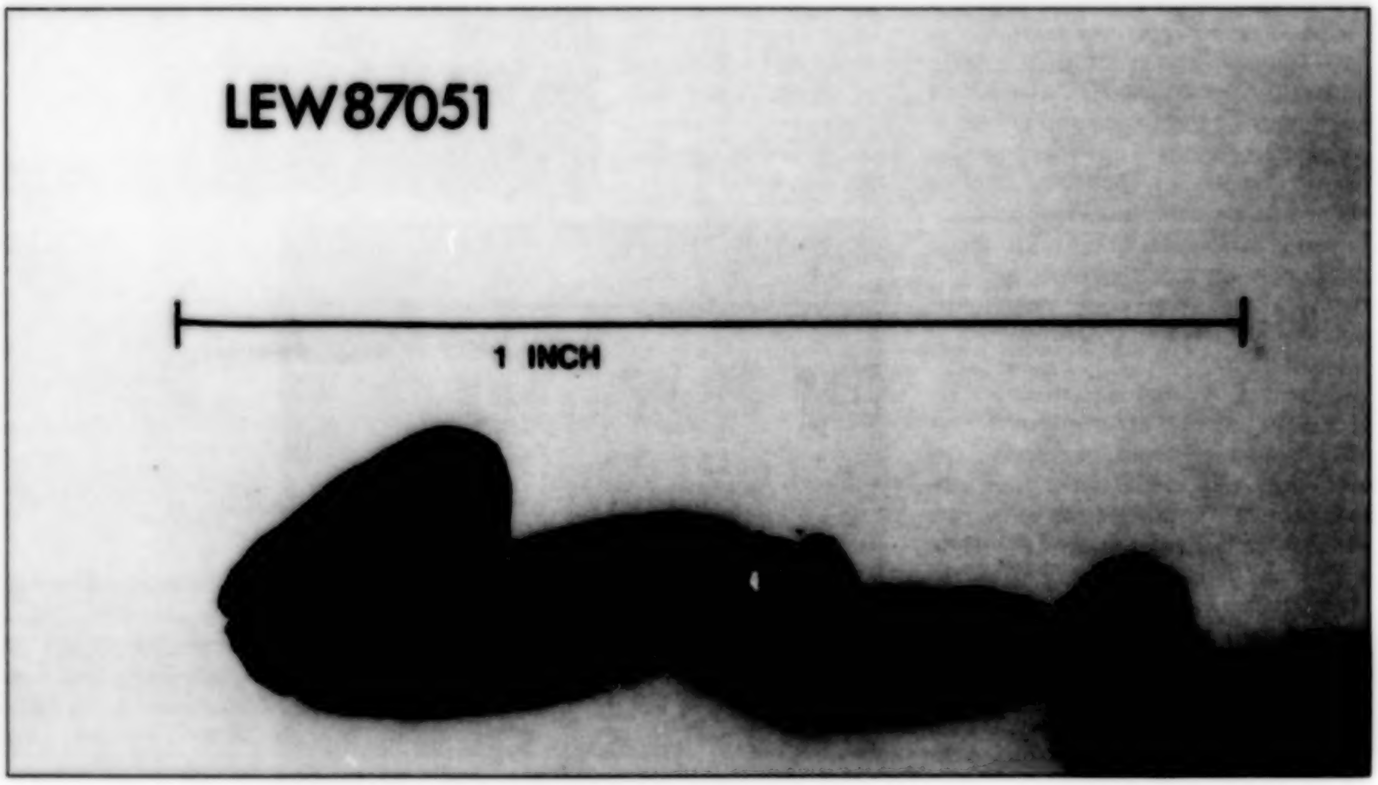


Fig. 2. The total recovered mass of Little LEW compared to a spatial scale of an inch. Analysis of such a small rock is only possible because of earlier work on lunar samples. Total weight is 1/50 of an ounce.

done to settle an argument between the two authors of this report. Mittlefehldt claimed that incomplete melting of Allende material would produce a silicate melt of the bulk elemental composition of an angrite plus residual, unmelted olivine. Jones argued that the melt produced would have the elemental composition of a eucrite.

We performed a series of experiments, all at a constant temperature of 1200° C (2200° F) but at different conditions of oxidation-reduction. The results were surprising. At the most reducing conditions, similar to those inferred for eucrites, the melts that formed had bulk elemental compositions like those of eucrites. At the most oxidizing conditions, similar to those inferred for angrites, the melts were similar in bulk composition to angrites. Except for the ratio of iron to magnesium, the melts produced in our most oxidizing experiments were extremely similar in bulk elemental composition to Lewis Cliff 86010. Additionally, under oxidizing conditions, the unmelted residue consisted of calcium-rich olivine like the olivine in Lewis Cliff 87051. Apparently by simply changing conditions of oxidation-reduction, it is possible to produce eucrite-like basalts and angrite-like basalts by melting on the same planet. Perhaps eucrites and angrites are related, and this relationship is the source of the isotopic similarities between the two types of meteorites.

The future. As we have emphasized, angrites are rare among meteorites on Earth. However, carbonaceous chondrites are not so rare, and several varieties are known. Based on our experiments, incomplete melting on parent bodies that consist of carbonaceous chondrite are just as likely to produce angritic basalts as eucritic basalts. Consequently, we suspect that angrites may not be as rare in the solar system as we once believed. When we eventually explore the asteroid belt and document the types of rocks we find there, we suspect that angrites will be prominent among the basalts found. The sooner our prediction can be tested, the better.

NITROGEN ON THE SURFACE OF THE MOON: CONNECTIONS WITH THE MANTLE OF THE EARTH AND WITH THE ANCIENT SUN

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Although the compositions of the Earth and the Moon are quite similar for many of the chemical elements, the volatile elements that on Earth make up the atmosphere, the hydrosphere and much of the biosphere were apparently lost from the Moon very early in its history. (Oxygen is present on the Moon only in solid chemical compounds such as silicates and oxides that make up lunar rocks.) One such element, abundantly familiar to us on Earth but essentially absent from the Moon, is nitrogen; typical lunar rocks contain less than one part per million. However, two types of lunar samples returned by the Apollo missions are found to contain modest amounts of nitrogen, and those occurrences of nitrogen on the Moon shed light on two quite different issues in cosmochemistry.

An exciting feature of the Apollo 17 mission was the discovery of a deposit of small beads of orange volcanic glass. These beads represent molten rock that solidified so quickly that, in many of the beads, crystals did not grow (Fig. 1). The beads apparently formed by a violent form of volcanism, known as lava fountaining, nearly 4 billion years ago. It has long been known that the surfaces of these glass beads are covered with a coating that is rich in volatile elements. In a recent study, the author, in collaboration with Kurt Marti and Jin S. Kim (University of California, San Diego) and Otto Eugster (on leave from the University of Bern, Switzerland), found that some, but not all, of these coatings contain small quantities of nitrogen. Apparently, the gas that drove the lava fountaining contained nitrogen exhaled from the lunar interior, and conditions during the volcanism occasionally permitted some of this nitrogen to be retained on the surfaces of the beads. Thus, the orange glass has given us our first detectable samples of indigenous lunar nitrogen; apparently the Moon was not completely degassed during accretion, but continued to lose its modest initial volatile inventory for at least some hundreds of millions of years after formation.

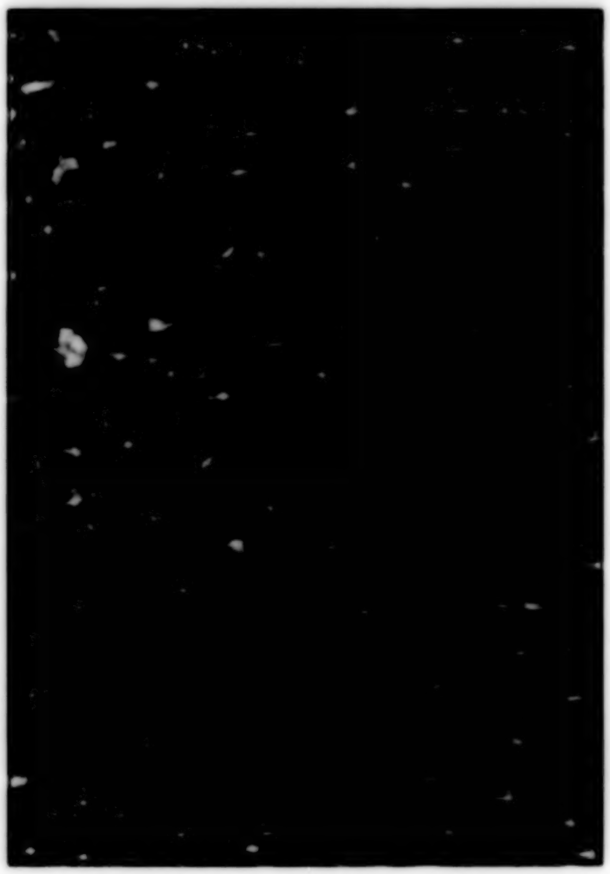

Fig. 1. Orange glass beads collected during the Apollo 17 mission. This material was formed by lava fountaining on the lunar surface nearly 4 billion years ago. Some of the beads are black, rather than orange, because of the presence of very small crystals that have subsequently grown in the glass. Width of field of view is a tenth of an inch. Photomicrograph by courtesy of Otto Eugster (University of Bern).

We cannot presently estimate how much nitrogen was incorporated into the Moon when it formed, but what is interesting is that we can determine the isotopic composition of that nitrogen. The isotopic composition we found is identical to that found by Colin Pillinger and his co-workers (The Open University, England) for nitrogen from the Loihi Seamount in the Pacific Ocean near Hawaii. This finding is significant because noble-gas studies have suggested that Loihi may be tapping a primitive region of the terrestrial mantle. Although the signature of mantle nitrogen is still tentative, these isotopic data reinforce the idea of a close genetic link between the Moon and the mantle of the Earth.

The other nitrogen-bearing material returned from the Moon consists of samples from the layer of finely pulverized rock that blankets most of the lunar surface, known as the regolith. It is well known that exposure to the space environment results in a progressive build-up of extralunar material in the regolith. Such material includes debris from meteorite impacts and ions

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implanted by the solar wind and solar flares. One of the elements whose concentration in the regolith increases progressively with increasing surface exposure is nitrogen. Clearly the great majority of these nitrogen atoms are of extralunar origin, but the actual source is controversial at this time. A major contribution is certainly from the solar wind, but we cannot rigorously rule out the possibility of a significant amount being supplied by some other source.

As with volcanic nitrogen, our interest in regolith nitrogen rests in its isotopic composition. Different regolith samples display wide variations in the ratio of the two stable nitrogen isotopes, whose masses equal 14 and 15 atomic mass units, respectively. Relative to the ratio found in the terrestrial atmosphere, these variations range from a 20% depletion of nitrogen-15, observed by Mark Thiemens (then at the University of Chicago, now at the University of California, San Diego) to a 20% enrichment in nitrogen-15, found by Richard Becker (University of Minnesota).

In 1975, I showed that this variation appeared to be related to the epoch during which the regolith samples were exposed on the lunar surface. Samples that were exposed a long time ago had systematically lower proportions of nitrogen-15 than those exposed more recently (Fig. 2). This observation was confirmed almost immediately by Richard Becker and Robert Clayton (University of Chicago). However, despite widespread acceptance of the validity of the effect, and early recognition that it constitutes one of the most remarkable scientific findings of the Apollo program, fifteen years later it still lacks a convincing explanation.

My original suggestion that the phenomenon represents a long-term increase in the proportion of nitrogen-15 in the solar wind has been criticized by Johannes Geiss and Peter Bochsler (University of Bern), because it is inconsistent with what is known about the structure and evolution of the Sun. An alternative theory, put forward by Geiss and Bochsler, suggests that in ancient samples, the solar wind, which they postulate to be enriched in nitrogen-15, was mixed with a hypothetical nonsolar nitrogen component poor in nitrogen-15. However, I have criticized this theory on the grounds that the relationship between nitrogen abundance and duration of surface exposure points toward a single, presumably solar, source for the regolith nitrogen, rather than the dual sources proposed by Geiss and Bochsler.

That is where the matter rests at present. The Apollo samples have presented us with an enigma whose eventual resolution must have profound implications for either lunar or, as I believe, solar evolution. In all likelihood, it remains for the next generation of lunar samples to supply that resolution.

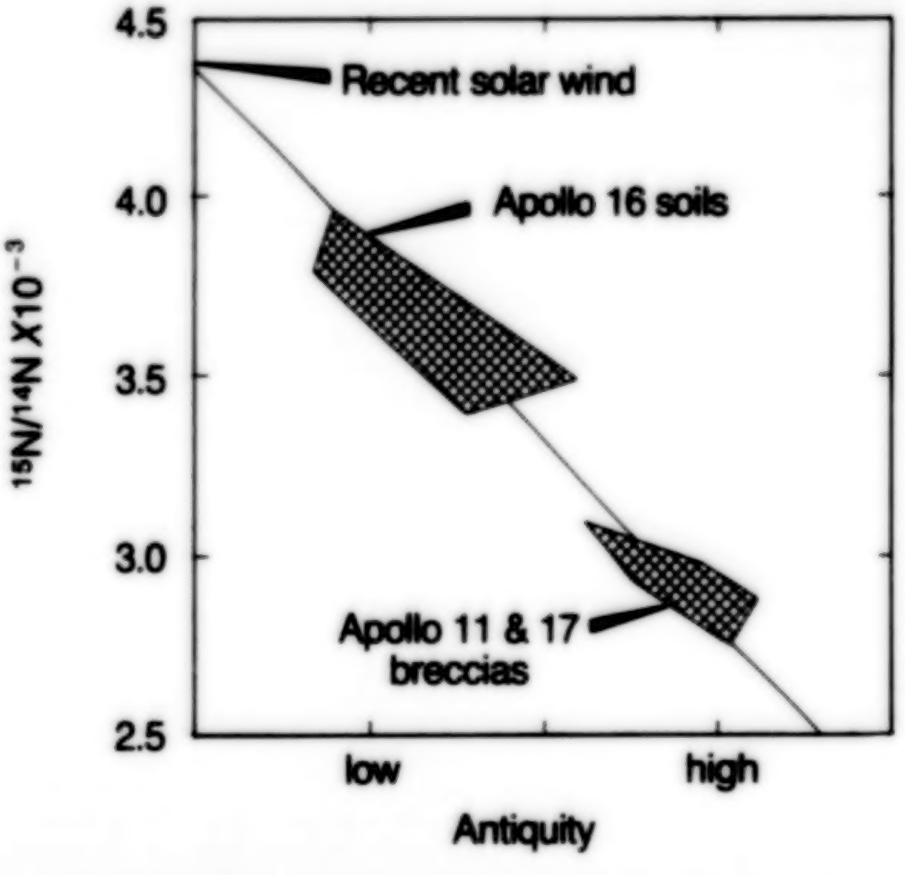

Fig. 2. Graphic representation of the long-term change in the isotopic composition of nitrogen in the lunar regolith. Antiquity is a measure of how long ago a sample was exposed on the lunar surface; on this plot, "high" antiquity probably corresponds to exposure roughly 3 billion years ago, but techniques for measuring antiquity are poorly calibrated at present. Note that the change amounts to a 50% increase in the proportion of nitrogen -15.

EMBEDDED MOONLETS: THE HOLY GRAIL OF PLANETARY RING RESEARCH

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Were Saturn's rings formed, perhaps even recently, by the catastrophic breakup of a stray comet or a satellite that drifted too close to the planet? Or are they retilled remnants of primordial disks, planetary counterparts to the nebula out of which our solar system itself formed? This, the most fundamental, question of ring origin may be settled by the discovery of large moonlets (radii of a few miles or more) within the rings since most models of accretion suggest that circumplanetary particles will have limited growth if the accumulation occurs near the planet while collisional fragmentation will leave some large objects intact.

Moonlets so small as to be unseen (or very difficult to see) by the Voyager cameras can still have a significant effect on the structure of planetary rings. Such ring denizens lie in a size range that is intermediate between typical ring particles (inches to yards) and the larger satellites that are visually observed to skirt the ring systems of the giant planets. For the rings of Saturn, these moonlets can be embedded amidst ring material, orbit in (and likely cause) a gap, or move in close proximity to an isolated ring (much as Prometheus and Pandora, the "shepherding" moons of Saturn's F ring, do). These bodies can then be discerned by seeking indirect evidence of their effect, usually expressed through their gravitational signatures.

The most notable example of detective work to find unseen embedded moonlets has concerned the neighborhood of Saturn's Encke gap, a 200 mile opening in the outer A ring. J.N. Cuzzi and J.D. Scargle (NASA Ames Research Center) and subsequently M.R. Showalter (Stanford University) and colleagues analyzed the gap's wavy edges, which oscillate by a mile or so, to identify the gravitational wakes of a small, previously unknown, moonlet in the gap. Showalter has recently discovered actual images of the Encke moonlet in a thorough perusal of the Voyager 2 imaging data.

Other researchers have suggested that gaps in Saturn's C ring and Cassini Division may have been pried apart by the gravitational torques induced by embedded moonlets. Small bodies have also been suggested to explain some of the bizarre behavior of Saturn's F ring (see Fig. 1), specifically the braids. J.J. Lissauer (State University of New York, Stony Brook) and J.S. Peale (University of California, Santa Barbara) have shown that, under certain conditions, and in concert with a larger perturber, a small body (about 3 miles in radius) can create braidlike features.

Our study of the F ring's structure (to be described in more detail below) indicates at least one additional moon close to the F ring from its gravitational imprint on the ring. Ongoing research by M.L. Cooke, P.D. Nicholson (Cornell University) and M.R. Showalter reveals that the 21 mile width of the Keeler gap near the outer perimeter of the A ring varies sinusoidally by ±15 - 20%; perhaps a moonlet or two is at work again. Some odd behavior in the Uranian and Neptunian rings may well be explained by the influence of small proximate bodies.

Since Voyager 1 first detailed the F ring's bizarre combination of kinks, clumps and braids in 1980, this outermost detached ring of Saturn has remained enigmatic, stubbornly refusing to yield its secrets. Figure 1 shows the F ring as a narrow (width of a few tens of miles) and slightly eccentric ring, lying some 2100 miles distant from the outer edge of the main ring system. The F ring is known to be variable, changing its appearance with time, position and viewing geometry. Initial Voyager reports noted that the various irregularities showed a range of wavelengths visible between 4,000 and 9,000 miles (corrected from an erroneous earlier value of 400 - 900 miles). The moons Prometheus and Pandora, which orbit interior and exterior to the ring, are thought to confine the ring (through the "shepherding" mechanism developed by P. Goldreich, Caltech, and S. Tremaine, Canadian Institute of Theoretical Astronomy), as well as to generate its kinks and clumps (M.R. Showalter and J.A. Burns). But, to date, no cogent models describing the ring's complete behavior have been developed.

Fig. 1. The F ring, as viewed by Voyager 1. In this image, the F ring's kinks, clumps and braids can be seen. The different colors correspond to a computer enhancement of the contrast in the image.

Following the initial effort of the Voyager team on quantifying the observed behavior of the ring, the only study has been that of J.B. Pollack and co-workers (NASA Ames Research Center) who are currently characterizing the photometric properties of the ring.

Our recent analysis of the F ring illustrates the utility of an indirect technique, namely the search for, and analysis of, periodicities in the ring. Brightness variations are a key diagnostic to narrow ring behavior because they can highlight the local perturbations of nearby bodies. Simple gravitational physics tells us that a nearby moon will induce a periodic perturbation on a narrow ring with a wavelength that is 3π times the difference in mean orbital separation of the perturber and the ring.

Thus, we hoped that by analyzing the striking azimuthal variations in the brightness of Saturn's F ring, we might quantitatively describe the various features and thereby identify their causes. Would we find a range of wavelengths as in the preliminary Voyager results, or only the signatures of Prometheus and Pandora? In our set of images, we observed two bright clumps of material moving at the same rate as the ring; these might be relatively large embedded bodies (moonlets), but unfortunately the image resolution is too low to allow an unequivocal identification of the clumps as moonlets. Figure 2 shows how integrated brightness (equivalent width) varies over 150° of the F ring's length.

In order to separate out any individual periodic signals in the data, we used a technique known as a Fast Fourier Transform. We were able to identify several distinct signals to be present, one of which is the gravitational signature of Prometheus, the larger of the two shepherds close to the F ring. Another signal can be explained as due to a previously undetected small satellite (radius of a few miles, on an eccentric orbit

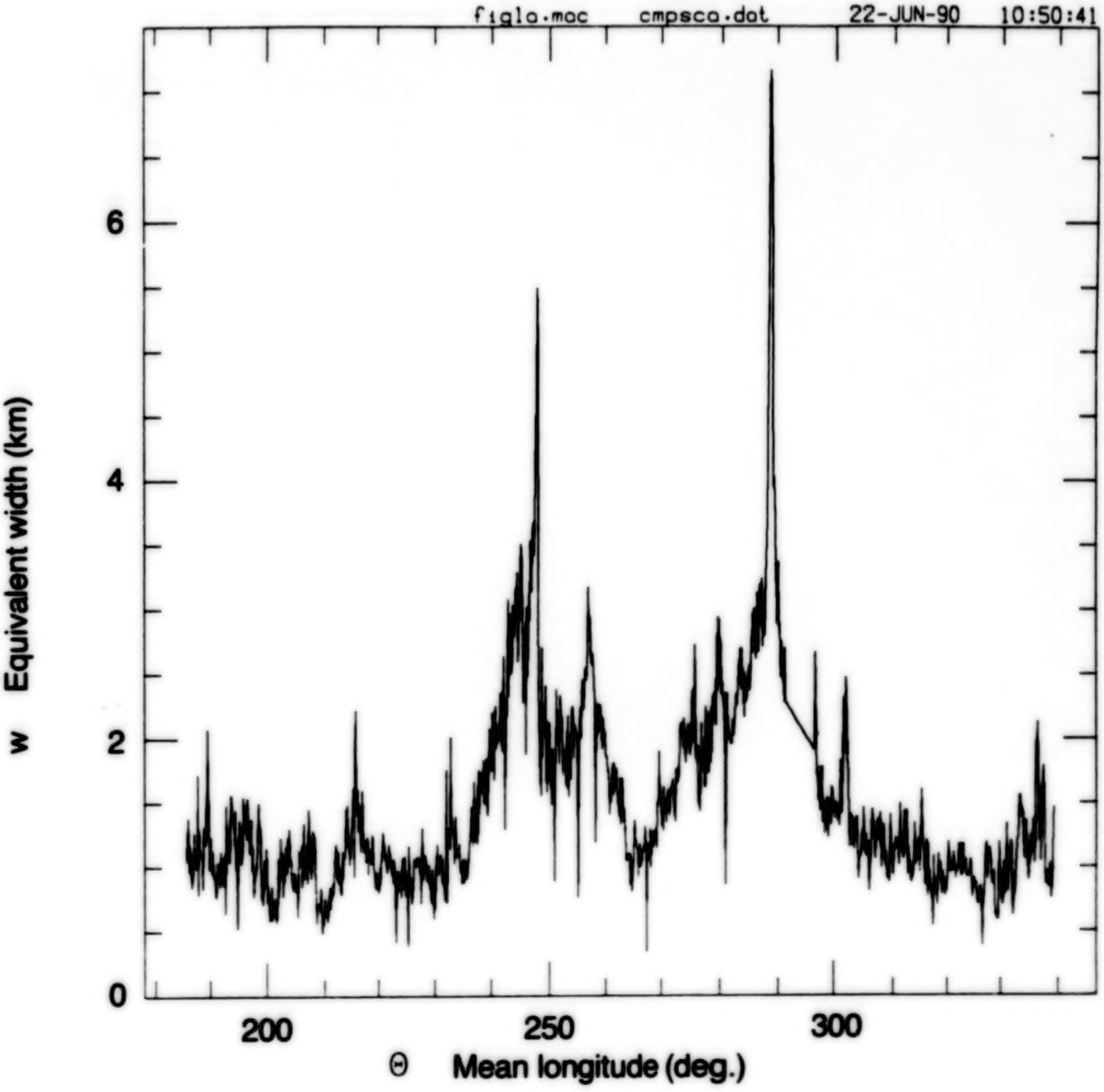

Fig. 2. A plot of equivalent width (1 km = 0.6 miles) versus longitude for our study of the F ring. The two larger peaks correspond to two bright clumps observable in the images. Note that it is very difficult to pick out periodicities in the data by eye, forcing us to employ Fourier analysis (as described in the text).

with a mean orbital separation of about 700 miles) orbiting near the F ring. The idea that a swarm of unseen satellites might be moving close to the F ring is not new: J.N. Cuzzi and J.A. Burns have argued that such objects might explain anomalous measurements of charged particle absorptions made by Pioneer 11.

Small bodies can have significant influence on planetary rings and may provide strong clues in favor of a catastrophic origin for the rings. A detailed understanding of their dynamics is very important in trying to interpret the data from Voyager. Given the long wait for additional imaging results on the Saturn system, these indirect tests help to better characterize planetary rings. Hopefully, the images provided by the Cassini mission to Saturn will be able to actually catch sight of more of these elusive moonlets.

THE NATURE OF THE ANCIENT SOLAR NEBULA: CLUES FROM ISOTOPIC STUDIES OF PRIMITIVE METEORITES

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Virtually all of the chemical elements in the universe, other than hydrogen and helium, are formed by nuclear reactions inside of stars. When those stars run out of fuel they "die". The death of a massive star, which may be 10 to 50 times the size of our Sun, is a spectacular event — it becomes a supernova star such as the recently observed supernova SN 1987A in the Large Magellenic Cloud. Within the course of only a few seconds the star collapses into itself, compressing its interior to exceedingly high densities and temperatures of several billion degrees are reached. This is followed by a cataclysmic explosion that expels a significant fraction of the stars' component matter (including the newly formed heavy elements) out into space.

The vast clouds of dust and gas that exist in the voids between stars contain matter accumulated from all the stars that have died since the galaxy was formed billions of years ago. Because those clouds in turn are the places where new stars are born, each new generation of stars starts out with a somewhat higher proportion of heavy elements than previous generations. Our solar system formed approximately 4 1/2 billion years ago out of such a cloud that contained many generations of reprocessed matter. Thus, the heavy elements (e.g. iron, silicon, oxygen, etc.) that make up most of the planets today already existed in the cloud at the time the solar system formed. It is likely that the "primordial cloud" was very heterogeneous, being a mixture of the expelled matter from many generations of stars that were born and died long before our solar system was born.

The solar system began to form when a small part of the cloud collapsed under its own gravitational weight, shrinking into a rotating disk (called the solar nebula) that was a turbulent and locally very hot place. Surprisingly, in spite of all the turbulence and violence involved in the making of the planets and the Sun, physical and chemical traces of this ancient interstellar cloud are preserved in small grains within rare meteorites known as carbonaceous chondrites. These grains consist of so-called "refractory" compounds that form only at very high temperatures; their closest analogues here on Earth are ceramics, such as the materials used to line the walls of blast furnaces. The carbonaceous chondrites are aggregates of the primitive solid matter that first lumped together to form the planets; they have been preserved in their original pristine state because they never became incorporated into planets large enough to be geologically active. In contrast, for example, all traces of the preplanetary dust from which our Earth formed have been obliterated by the eons of melting, recrystallizing, mountain-building and weathering that shaped our planet.

Since 1973, it has been known that the "refractory" particles within carbonaceous chondrite meteorites possess in their isotopic composition a distinctive signature that can be directly traced back to one or more dying stars. In other words, atomic species of a particular chemical element (i.e. its isotopes) show small but significant differences from their respective relative abundances found today in average solar system material such as on the Earth or Moon.

The refractory particles are not, themselves, physical remnants of the dust from interstellar space. They formed by a complex series of vaporization and recondensation processes within the solar nebula, and their isotopic compositions, therefore, reflect the compositions of the nebular gas. As more and more refractory grains have been analyzed, it has become clear that a considerable variation exists in their isotopic composition. This has been attributed to small scale and local heterogeneities within the nebular gas.

Of these "refractory" particles, which are themselves relatively rare, an even more exclusive subgroup exists. Six grains have been found so far out of the hundreds of refractory grains analyzed that are particularly exotic in their isotopic compositions even relative to the other refractory grains. These objects were thought to have formed in very small pockets of isotopically unusual gas within the solar nebula. The dust was presumably interstellar dust of highly unusual isotopic

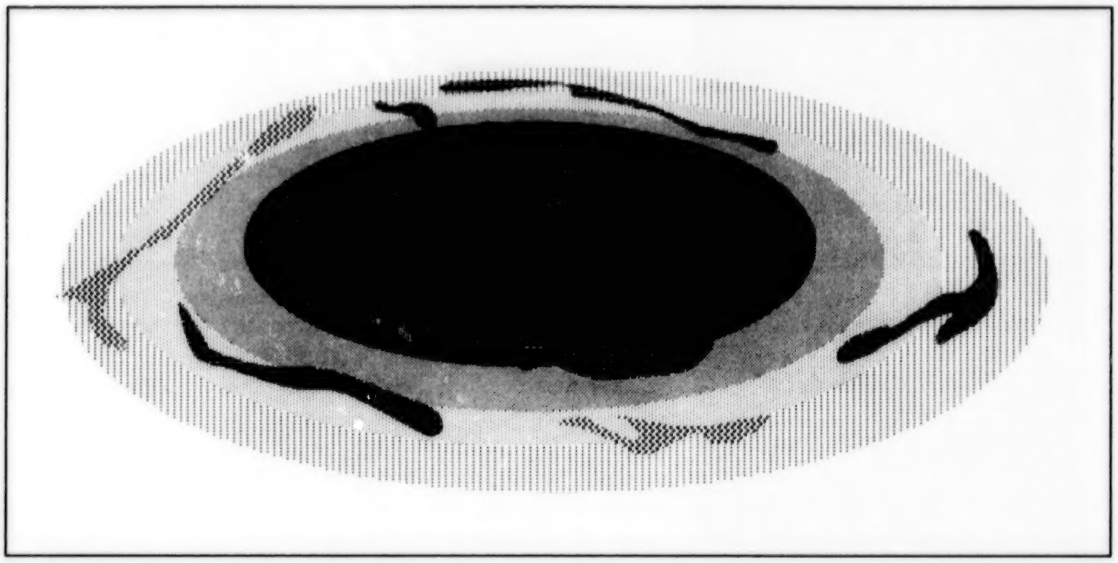

Fig. 1. Conceptual view of the solar nebula 4 1/2 billion years ago, before formation of the Sun and planets, showing small "cloudlets" that contain isotopically different material from the rest of the gas and dust. The view is oblique, from somewhat above the plane of the disk.

composition.

Recently, however, R. Loss and G. W. Lugmair (University of California, San Diego), G. MacPherson (Smithsonian Institution) and A. Davis (University of Chicago) found and analyzed a new grain that has isotopic properties nearly identical with another particle found years ago in a different meteorite; yet, it is mineralogically different and has experienced a very different evolutionary history. This is a very important finding because this apparent "coincidence" for the first time provides evidence that some of the isotopically distinct regions of the solar nebula were not as small and presumably short-lived as previously thought. The only reasonable way for such regions to be preserved over long periods of time (possibly thousands or even tens of thousands of years) is if the isotopically unusual material was largely contained in dust rather than gas. This implies that the surrounding nebular gas was not hot, a concept that has been increasingly advocated over the past ten years and which runs counter to the older traditional model in which most or all of the original interstellar dust was vaporized within a globally hot nebula.

It may seem odd, even ironic, that much of our knowledge about the earliest formation of our solar system is derived from studies of small grains that are only fractions of an inch or so in size. Yet, such studies over the past 10-20 years have not only greatly enlightened our view of how the solar system originated; these results help constrain astrophysical models of how elements are made within massive stars and, even, how planetary systems are formed from interstellar clouds. In addition, the need for ever increasing refinement in our methods in order to unlock nature's secrets has greatly advanced many aspects of our measurement technology.

REMOTE SENSING OF THE ORIENTALE BASIN OF THE MOON: SETTING THE STAGE FOR GALILEO'S FIRST ENCOUNTER WITH THE MOON

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In early December 1990, the Galileo spacecraft will fly through the Earth-Moon system in a maneuver that will use the gravity of the Earth to increase Galileo's velocity and shorten its travel time to Jupiter. During this encounter, Galileo's powerful complement of remote sensing instruments, designed to study the jovian system, will be trained on the Earth and Moon. Galileo's trajectory will allow it to obtain remote sensing data for portions of the farside of the Moon, which is hidden from the Earth.

Until now, these types of data have been confined to the frontside collected by large observatories on Earth. A primary goal of Galileo's Moon encounter will be observations of a large impact structure centered on the western limb of the Moon such that only about half the structure is visible from the Earth. This feature is a huge multiringed impact crater known as the Orientale Basin.

Orientale formed by impact into the Moon of a very large asteroid, tens of miles in diameter, which left an impact structure consisting of concentric rings hundreds of miles in diameter. There are many such impact basins on the Moon, and elsewhere in the solar system, but Orientale is the youngest and best preserved of the lunar multiringed basins and one of the best preserved such structures in the solar system. Study of this basin is important to planetary science in general: large impacts early in the history of the solar system were a major force in forming terrains and affecting the evolution of planetary crusts. Lunar scientists are particularly interested in Orientale because they believe that these huge impacts excavate material from deep in the Moon, allowing study of the composition of the lunar interior. Indeed, impact craters have been called the "drill cores of the Moon," and Orientale was one of the deepest probes.

While Galileo will be well positioned to view all of Orientale, the location of the basin on the limb of the Moon allows portions of the basin to be studied from Earth. Indeed, Orientale was first identified as a multiringed structure by astronomers at the University of Arizona who took photographs of the lunar limb, projected these photographs on a large white globe and photographed the globe from the side. The eastern portions of the rings of Orientale were beautifully displayed by this technique. At times, Earth-bound astronomers can see a bit more than half of the basin. The Moon wobbles about its axis as viewed from the Earth, a phenomenon known as libration, bringing Orientale about 5° around the limb from its average position. So, from the Earth, astronomers are able to study slightly more than half the basin with all the remote sensing instruments at their disposal.

In anticipation of Galileo's encounter, we at the University of Hawaii (myself, B. Ray Hawke, and G. Jeffrey Taylor), and Paul Spudis of the Lunar and Planetary Institute in Houston, embarked on an extensive observing campaign to obtain data for the Orientale basin. Our three main reasons for this endeavor were: first, study of this region addresses fundamental questions about the evolution of the Moon's crust and about cratering dynamics; second, the results provide an initial scientific context for the Galileo remote sensing science teams, similar to the way data from the Voyager missions will be used as a guide for the Galileo's encounter with the Jupiter system; and third, data collected with instruments similar to those on Galileo and of the same areas on the Moon help in the calibration of Galileo's instruments.

During a favorable libration in August 1990, we used an infrared spectrometer to collect spectra (measurements of the reflectance or brightness of small portions of the lunar surface in 200 sequential infrared colors from which mineralogical content can be derived) similar to that of Galileo's Near Infrared Mapping Spectrometer (NIMS), and an electronic camera to collect images at many wavelengths similar to Galileo's Solid State Imaging (SSI) experiment. We also used an infrared camera which has no equivalent on Galileo but obtains SSI type imaging data at NIMS infrared wavelengths. The experiment, conducted simultaneously on

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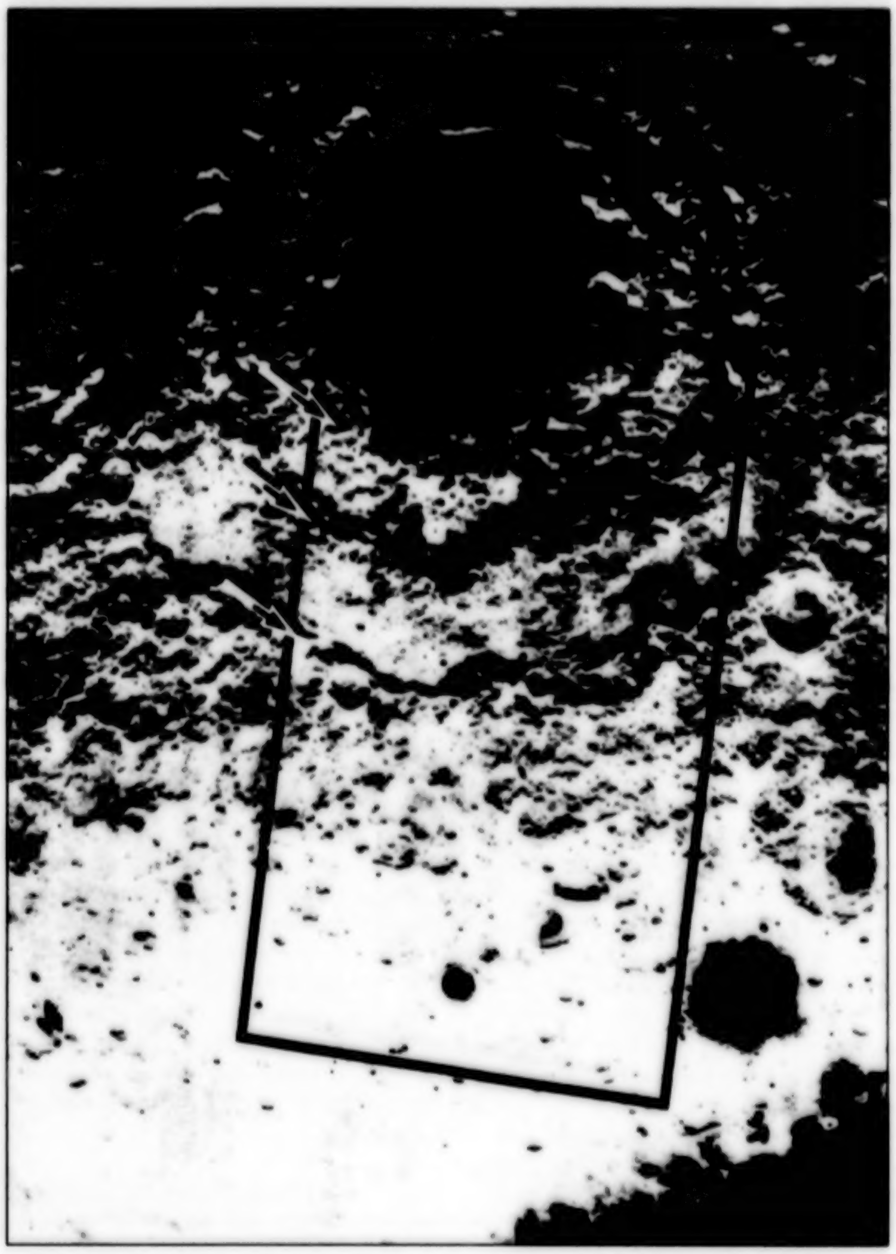

Fig. 1. Lunar Orbiter photograph of the Orientale Basin. The black box indicates the position of the accompanying color image. The three arrows indicate the major basin rings. From left to right they are: the Inner Rook Ring, the Outer Rook Ring, and the Cordillara Ring.

three telescopes at Mauna Kea Observatory on the island of Hawaii, was hugely successful, resulting in complete coverage of the visible portions of Orientale with the imaging experiments, and more than 40 high-quality spectra of small areas.

In 1984, Paul Spudis, then at the USGS Flagstaff, B. Ray Hawke and I did a study of rock types at Orientale. We used elemental measurements obtained by gamma-ray spectrometers on Apollos 15 and 16 of the northern portion of the ejecta blanket of Orientale. With our near-IR spectrometer on the University of Hawaii 88-inch telescope at Mauna Kea Observatory, we also obtained mineralogical measurements of a few small areas within the basin rings. Though the elemental measurements were fairly far from the basin center and the spectral measurements were sparse, we concluded that only two rock types were present in the Orientale region: anorthosite and noritic anorthosite.

These rock types have an extremely high abundance of the mineral anorthite, a plagioclase feldspar whose presence suggests an unusual origin for these rocks. Ordinary igneous processes should produce basaltic rocks with roughly equal amounts of feldspar and a more iron-rich mineral called pyroxene. Basalts are found in the maria of the Moon, the dark "seas" visible from Earth. Feldspar-rich rock types are found at the Apollo 16 landing site deep within the central highlands of the Moon, and are also found at other landing sites.

The presence of rocks of almost pure feldspar at some of the landing sites led to a model for lunar crustal origin known as the "magma ocean" hypothesis. This model holds that the excess feldspar found on the lunar surface formed from a huge ocean of molten rock, hundreds of kilometers deep, which may have enveloped the entire Moon. As the molten rock cooled, feldspar began to crystallize, and as this mineral is lighter than the liquid it floated to the top, eventually producing a feldspar rich crust. Proponents of competing models claim that the Apollo 16 landing site is atypical, reflecting local, not global, compositions. However, our early observations showed that the anorthosite-rich nature of the crust is also a feature of that region, supporting the "magma ocean" model.

Our recent data greatly improves our coverage of the Orientale region, and is beginning to reveal interesting distributions of the rocks we identified previously. First, the extensive spectral sampling of the ejecta of Orientale has shown that it is either identical to or even slightly richer in feldspar than the Apollo 16

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COLOR PHOTOGRAPH

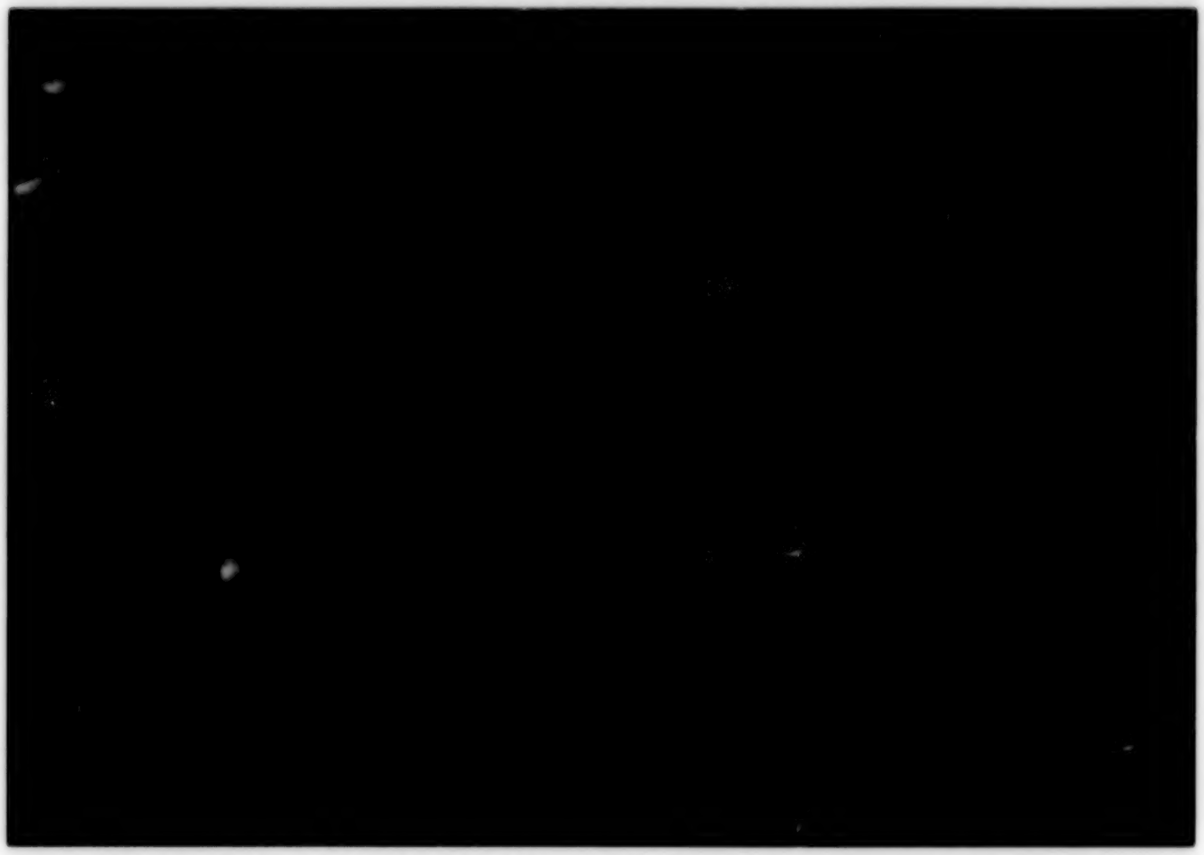

Fig. 2. Color enhanced image of a portion of the Orientale region shown in the accompanying image. Because the image was obtained near full Moon, shadows which make craters and other features so apparent in the accompanying image are absent. Their absence simplifies the interpretation of the color information. The position of the three major basin rings are shown with arrows. Bright yellow colors indicate the presence of anorthosite; blue and blue-white areas are more pyroxene rich. Mare basalt deposits appear black.

landing site. The vast region surrounding Orientale is confirmed to be highly anorthositic. This supports our earlier conclusions and shows that the types of rocks found at Apollo 16 are present in large amounts in at least one other part of the Moon.

The imaging experiment, the first product of which is shown here, displays the distribution of material that is essentially pure anorthite versus that which contains measurable amounts of pyroxene. In the color image, the bright areas with yellow hues are pure anorthite (less than about 2% pyroxene), and the bright areas with blue hues have substantial amounts of pyroxene (though the spectral data show that these areas are still anorthite-rich). It is apparent that the innermost major ring of Orientale is composed of pure anorthite, whereas the next rings show the presence of pyroxene-bearing material. (At this time, our imaging analysis is confined to bright features because flat-lying areas accumulate dark glass from fusion of soil by micrometeorite impacts. This dark glass partly masks the spectral character of the lunar soil and has other subtle effects on the images. Moonquakes and meteorite impacts continuously expose material free of this glass on steep mountain slopes and in crater walls through downslope movement of material, so we focus this type of analysis on bright, glass-free features.) Other small bright features with yellow and blue hues occur well away from the outer main ring of Orientale, suggesting that both pure anorthite areas and pyroxene bearing areas exist in these areas.

Pure anorthite in the inner main ring of Orientale is puzzling. According to the most favored cratering model, the material in this ring has rebounded from deep in the Moon, perhaps from about 30 miles depth. Thus the material of alleged deepest origin is free of pyroxene. However, according to the magma ocean hypothesis, the shallowest material should be the richest in feldspar. This observation seems to suggest that either our favored cratering model is incorrect and the inner ring does not represent the deepest material, or that the simplest version of the magma ocean model may not be correct and that the upper part of the crust has evolved, perhaps through intrusion or extrusion of more mafic magmas, since the time of formation of the magma ocean and solidification of the original crust.

Our preliminary work shows that the Orientale region is generally similar to the Apollo 16 landing site. However, the occurrence and distribution of the rock types cause problems for some of the theories that now form the basis of our understanding of the lunar crust. We anticipate that Galileo will be able to extend this work beyond the lunar limb to the western portion of Orientale and to other parts of the lunar farside, perhaps to shed further light on these puzzles.

GLOBAL COLOR AND ALBEDO VARIATIONS ON TRITON

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The Voyager 2 encounter with the Neptune system in August of 1989 revealed an active surface and atmosphere on Triton. Earth-based observations had indicated that nitrogen and methane occur on Triton's surface, and Voyager ultraviolet spectrometer measurements showed that the atmosphere is dominated by nitrogen. The "subsolar latitude" (latitude of points lying directly beneath the Sun) on Triton varies seasonally by as much as ±50°, so large seasonal variations in Triton's atmospheric pressure and surface frost distribution are expected. The subsolar latitude crossed the equator and passed into the southern hemisphere in 1930, reaching a latitude of -45° by 1989.

Pre-Voyager models predicted that, by 1989, most of Triton's northern hemisphere would be covered by bright frost, that frost-free areas near the equator would be covered by dark red organic compounds, and that the south polar cap (SPC) would extend to about -35° latitude. Contrary to these predictions, Voyager images show that the SPC covers nearly the entire southern hemisphere and has a bright fringe around its margin and that the illuminated part of the northern hemisphere (as far as +45° latitude) is relatively dark and reddish.

Temporal color changes on Triton also differ from pre-Voyager expectations. By 1989, Triton had become substantially less red than it was in 1977. The reddest area observed by Voyager is much less red than the 1977 full disk color. N_2 (molecular nitrogen) is very mobile at Triton's surface temperature of -390° F, so it seems plausible that N_2 transport, combined with the 12° change in subsolar and sub-Earth latitude between 1977 and 1989, could account for this large color change. However, this change was opposite that expected from "sublimation" (transition directly from solid to vapor state) and recession of the SPC.

In the work described here, variations in color and "albedo" (percentage of incoming radiation that is reflected) over most of the illuminated part of Triton were mapped (Fig. 1) from full-disk color sequences acquired by Voyager 2 at low phase angles and at various rotational longitudes. These mosaics and the photometric-function model were used to produce an albedo map for thermal-balance models in collaboration with John Stansberry and colleagues at the University of Arizona. With these datasets and models, some partial explanations for the observed large-scale color and albedo patterns and the 1977-1989 color change are suggested.

The Polar Stereographic projection (Fig. 1E) shows Triton's entire SPC, bright fringe, and north-trending rays. The fringe and rays (blue or blue-green in the false-color images) appear diffusely deposited over both the SPC and the relatively dark areas to the north. The polar-cap margin is highly irregular, but it has an overall bilateral symmetry centered on the sub-Neptune longitude (0°). Within about 45° of 0° longitude, the SPC extends up to about +10° latitude, whereas at other longitudes its margin ranges in latitude from the equator to -30° and shows a pattern of "scallops," each about 10° to 30° wide. Segments of the margin are straight or form polygons, and they transect morphologic units such as the "cantaloupe terrain"; these relations are reminiscent of the north polar ice cap of Mars.

The bright fringe and rays are closely associated with the cap's margin, irrespective of its latitude. The bright rays extend north-northeast for hundreds of miles, emanating preferentially from the points of the scalloped margin. There is a smooth transition in color and albedo from the bright fringe to the rays, which become fainter to the north. Spectral properties of the fringe are consistent with those of fresh frost or snow. These relations suggest that the rays consist of fine-grained frost or snow that was redistributed by northward winds associated with Triton's rotation.

The large size of the SPC was not predicted by pre-Voyager models of volatile transport, in which the summer cap margin was predicted to recede as the subsolar latitude increased. However, thermal-balance models have shown that bright parts of the SPC within about 30° of

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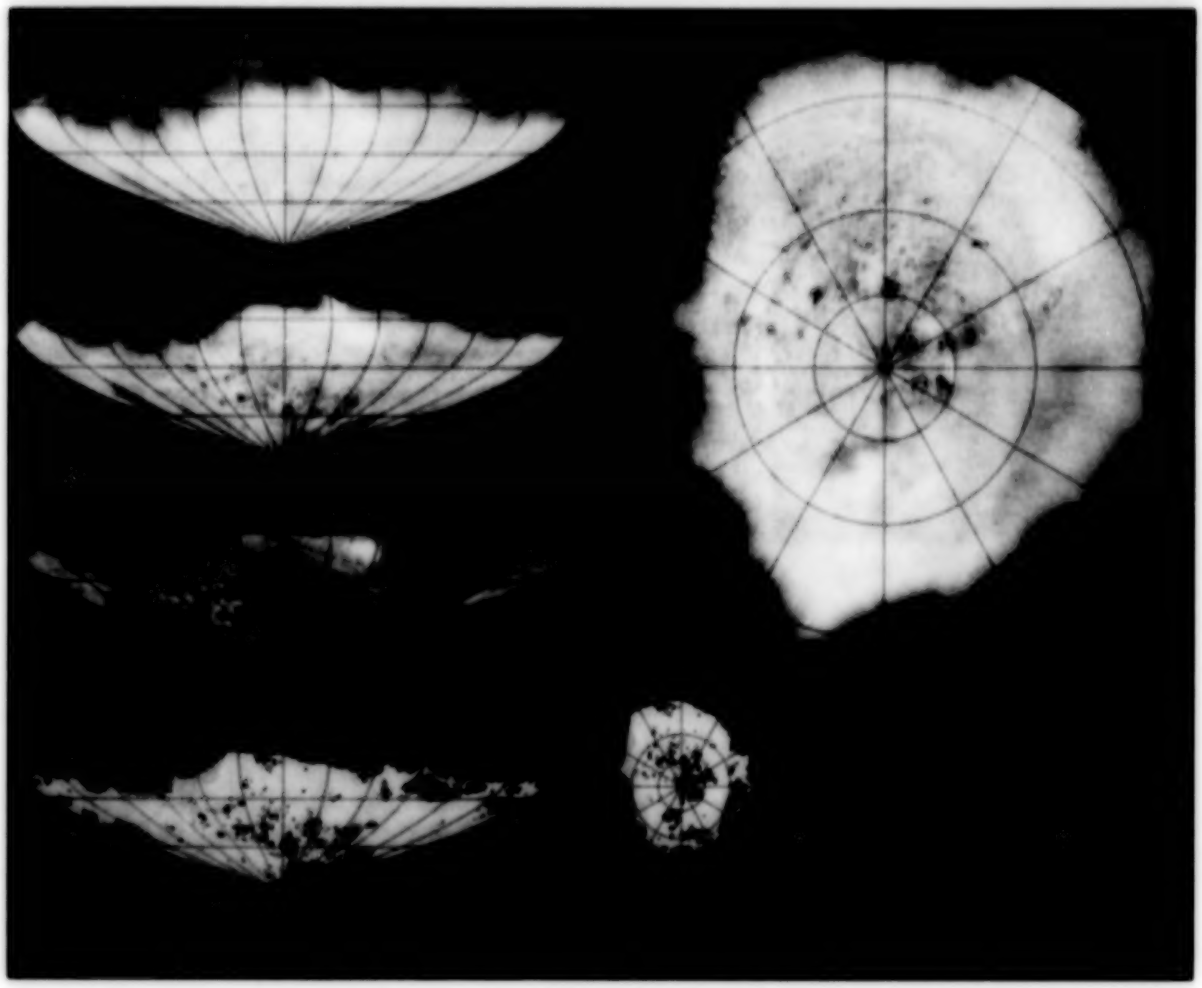

Fig. 1. Color mosaics and spectral-unit maps. A is approximately natural color (orange, green, and blue mosaics); B and E consist of green, violet, and uv mosaics with linear contrast stretches applied to each; and C consists of green, violet and uv mosaics with a decorrelation stretch to maximize color separation. D and F are unit maps; unit colors are (1) red, (2) green, (3) yellow, (4) purple, (5) buff, and (6) blue. A-D are Sinusoidal Equal-Area projections centered on 0° longitude; E-F are Polar Stereographic projections centered on the south pole with 0° longitude at top. Grids indicate 30° intervals of latitude and longitude.

the equator should be experiencing net condensation under current conditions, and that areas of average cap albedo in this zone and brighter areas as far south as -60° are nearly stable, with low sublimation rates.

On Mars, the caps recede at their margins while the subsolar point remains equatorward of these margins. On Triton, the subsolar latitude may have moved toward the south pole faster than the margin could retreat given the volatile inventory and the timescale for moving it. At present, maximum insolation occurs well within the cap, such that the margin has ceased to retreat and may actually advance.

The color mosaics have been classified into six major spectral units (Fig. 1D, F). Unit 5 (buff) corresponds to the major part of the SPC, which probably includes both CH_4 (methane) and N_2 frost or ice to account for Earth-based spectral observations. Unit 5 is interpreted as delineating the residual SPC, but significant areal coverage by seasonal N_2 frost is needed to support the observed atmospheric pressure (~14-19μb). Unit 2 (green) consists of the "dark" patches and streaks on the SPC, which are slightly darker and redder than unit 5. The dark streaks most likely result from fallout of dark materials erupted in recent "geysers" discovered by Voyager scientists. Unit 6 (blue) is the bright fringe along the northern margin of the SPC. It probably consists of very fresh N_2 frost, because its high albedo and northern position will cause N_2 deposition. Unit 1 (red) corresponds to the relatively dark and reddish northern equatorial region of the satellite, north of the SPC. Thermal-balance calculations indicate that N_2 frost should condense over regions north of about +15° latitude, but the reddish color and low albedo of this region are not expected for fresh frost, and the composition of this unit is not understood. Unit 4 (purple) is transitional to units 1 and 6 and is characterized by the north-trending rays. Unit 3 (yellow) is a bright but relatively reddish unit. It occurs in the only gap in the bright fringe, and reddish material, such as organic matter from atmospheric hazes or active geysers, may be superposed on the bright fringe.

The 1977-1989 change in Triton's color can be explained by the deposition of bright and spectrally neutral frost on reddish materials such as irradiated methane clathrate. Deposition of frost in the northern hemisphere is expected in the current southern summer, so in a first attempt to model the color change, it was assumed that units 1 and 4 (23% of the projected disk), had a much redder spectrum in 1977. The calculated red spectrum for these areas that is needed to match the 1977 spectrum has negative albedos in the violet and uv, a nonphysical result. The additional step of eliminating the bright fringe (i.e., giving unit 6 the spectrum of unit 5) is also insufficient to match the 1977 spectrum. This suggests that frost has been deposited during the past 12 years not only in the northern hemisphere, but also over much of the southern hemisphere (on the residual SPC).

The change in Triton's color from 1977 to 1989 was sudden relative to Triton's 165-year seasonal cycle; this suggests that some sort of triggering mechanism and instability is involved. Perhaps the triggering mechanism was the initiation of active venting and the instability involves the effect of albedo on the sublimation rate. This question and many others concerning Triton's volatile behavior will be addressed in future studies.

EUROPIUM IN MARE BASALTS: SUPPORT FOR A GLOBAL LUNAR MAGMA OCEAN?

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The lunar magma-ocean hypothesis: The orthodox view. One dramatic scientific result of the Apollo program was the discovery of evidence suggesting that the early Moon was largely or completely molten. According to this hypothesis, the Moon's surface once consisted of an incandescent "ocean" of red-hot silicate melt, or magma. As this magma ocean cooled, minerals began to crystallize, much like ice forming in a freezing pond.

Some minerals, such as the calcium-aluminum silicate called plagioclase, were lighter than the magma. Like ice, these minerals floated to the surface to form "rockbergs" which eventually coalesced to form the lunar crust. Other minerals, such as the iron-magnesium silicates olivine and pyroxene, were denser than the magma. Like stones dropped into a pond, these minerals sank to the bottom of the magma ocean to accumulate in a thick layer, called the lunar mantle. These basic elements of the magma-ocean hypothesis are illustrated schematically in Figure 1.

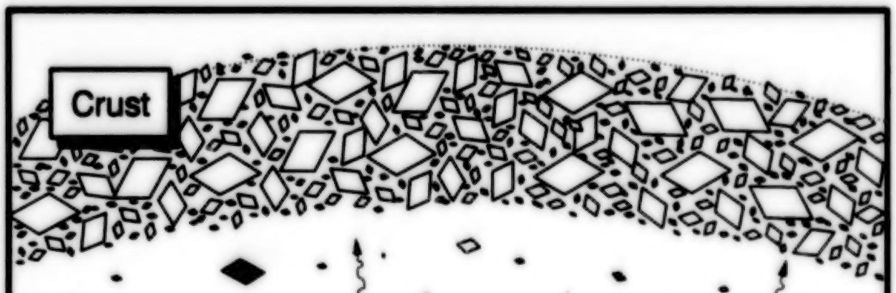

Fig. 1. Schematic cross-section of the early Moon, illustrating basic aspects of the magma-ocean hypothesis. According to this hypothesis, the outer part of the early Moon consisted of a global "ocean" of red-hot silicate melt, or magma. As this magma cooled, minerals crystallized. Light minerals, especially plagioclase, floated to the top of the ocean to form the lunar crust, while dense minerals such as olivine and pyroxene sank to the bottom of the ocean to form the mantle. This cartoon is based on a drawing by Dave Walker (Lamont-Doherty Geological Observatory).

According to the hypothesis, later reheating and melting of the olivine and pyroxene in the lunar mantle produced iron-rich molten lavas. These lavas migrated upward and erupted through huge systems of fissures to flood low lying areas of the lunar surface. The lavas solidified to form broad dark plains over much of the nearside of the Moon, and these plains are the dark areas on the face of the Moon visible to us with the naked eye. They are called "maria", or seas, and their rocks are called "mare" basalts.

Rare-earth elements: Tools for testing the hypothesis. Support for the magma-ocean hypothesis comes from the seemingly arcane study of a group of chemically related elements called the rare-earth elements. These elements have two felicitous properties which make them diagnostic of certain geochemical processes.

First, their chemical behavior is so similar that most of the rare earths are not easily separated from one another by common geological and geochemical processes. Hence, their relative abundances in most materials are similar to those in primitive, unprocessed solar system material (such as chondritic meteorites). Second, there is a fortunate exception to the above generalization. The chemical behavior of one of these elements, europium (Eu), is under certain conditions different from the others, so that the abundance of europium in a rock or mineral relative to the other rare-earth elements holds clues to the processes that have affected that rock or mineral.

Crystallization of plagioclase from a magma is one process that greatly affects relative europium abundances. Plagioclase has a much greater affinity for europium than for the other rare-earth elements. Hence, crystallization of plagioclase from a magma will preferentially extract europium from that magma; the plagioclase will be enriched in europium relative to the other rare earths and the magma will be depleted.

Until recently, most geochemists believed that the minerals olivine and pyroxene would not preferentially exclude or include europium, so that crystallization of these minerals should not change the abundance of europium relative to the other rare earths in a crystallizing magma. Minerals lacking a strong affinity for or aversion to europium will have the same ratio of europium to the other rare earths as the magma from which those minerals crystallized. Moreover, any lava formed by the subsequent remelting of such minerals will inherit the ratio that typified the original magma.

This understanding of the behavior of the rare earths during crystallization of a magma is one of the cornerstones of the lunar magma-ocean

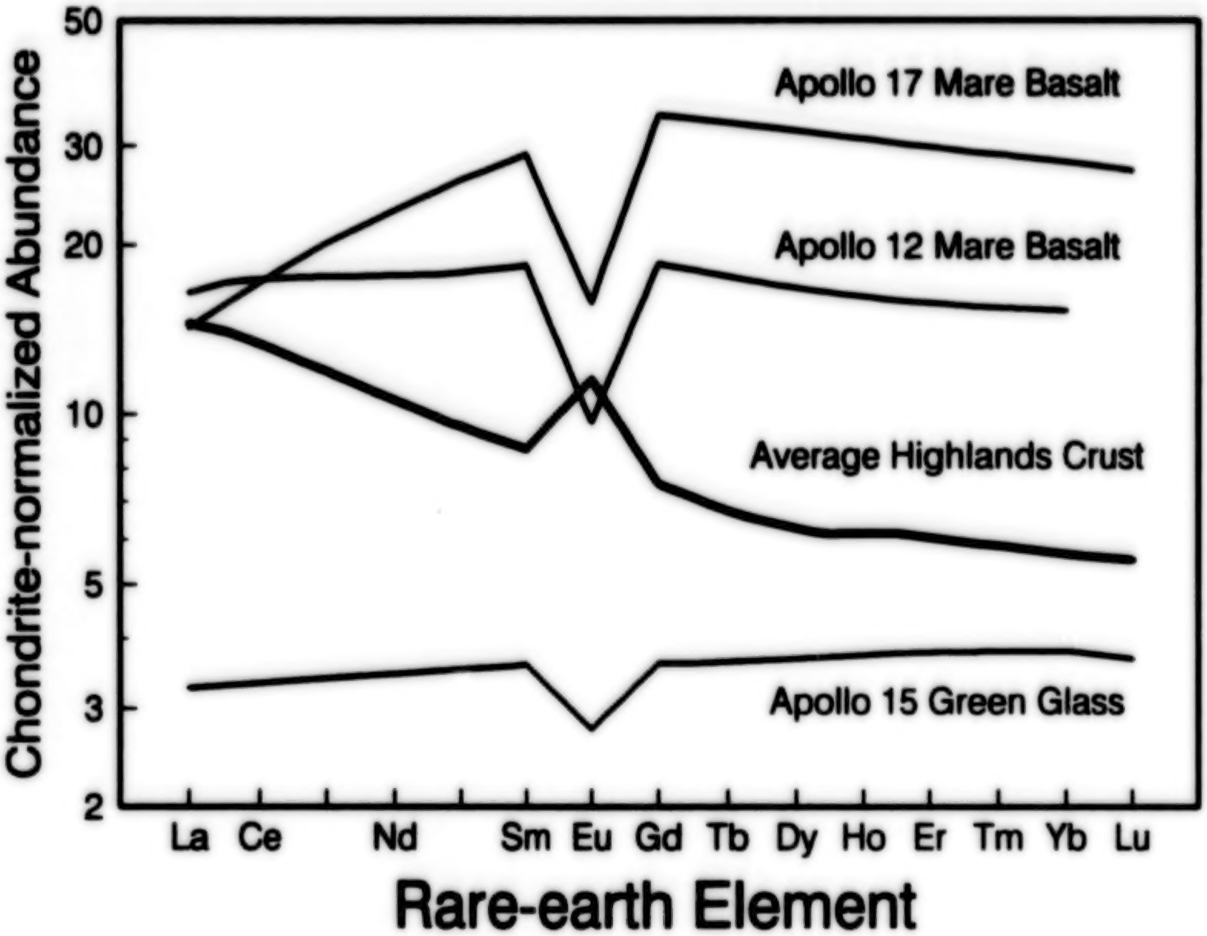

Fig. 2. Rare-earth element abundances in three lunar mare basalts and in the average lunar highlands crust composition proposed by Ross Taylor (Australian National University). (Apollo 15 Green Glass is an especially primitive form of mare basaltic lava, found as small glass beads which scientists think formed by rapid cooling of lava droplets in incandescent "fire fountains".) Abundances of rare earths have been normalized to chondritic meteorites to produce "smooth" patterns for elements other than europium. The rare earths are plotted in order of increasing atomic number, because, except for europium, their chemical properties correlate smoothly with atomic number. The "bump" at europium (Eu) in the highlands crust abundance pattern is a europium enrichment caused by the accumulation of plagioclase through flotation during the formation of the lunar crust. This bump is matched by a complementary "dip" or europium depletion for the mare basalt samples. The traditional interpretation of this depletion is that it was inherited from the depletion of europium in the magma ocean due to crystallization and flotation of plagioclase to form the highlands crust. These complementary europium enrichments and depletions are one of the major lines of evidence supporting the magma-ocean hypothesis.

theory. The mare-basalt lavas were depleted in europium relative to the other rare earths. Experimental melting studies have established that these lavas formed by melting of olivine and pyroxene in the lunar mantle. Thus, the olivine and pyroxene in the lunar mantle must also be depleted in europium.

How did this depletion arise? According to the magma-ocean theory, it arose because crystallization of plagioclase in a magma ocean and flotation of this plagioclase into the crust forming at the top of the ocean removed europium from the magma and enriched it in the crust. The olivine and pyroxene of the lunar mantle subsequently crystallized from the europium-depleted magma, inheriting its depletion. This depletion was in turn passed along to the mare-basalt lavas formed by later remelting of the olivine and pyroxene.

This scenario is particularly appealing because it ties together the rare-earth abundances in mare basalts and in the rocks of the lunar crust in a single, simple theory. Thus the complementary depletion of europium in the mare basalts and enrichment in the rocks of the lunar crust (Fig. 2) was one of the fundamental observations that led to the magma-ocean hypothesis. This hypothesis is currently widely accepted among lunar scientists. Occasionally, however, a few heretics have questioned the theory, and some have proposed that early melting on the Moon was never "oceanic" in scale.

A heretical view: Rare-earth evidence questioned. The most recent outbreak of such heresy came from Chip Shearer and Jim Papike (South Dakota School of Mines). Using a new, ultrasensitive technique for measuring rare-earth element abundances in individual mineral grains, they noted that pyroxene showed a surprisingly large depletion in europium relative to the other rare earths. They suggested that this depletion was due to a previously unrecognized tendency of pyroxene to exclude europium, and further suggested that the europium depletion of the mare-basalt lavas was not inherited from a europium-depleted magma ocean after all, but instead was produced as a result of this intrinsic property of pyroxene.

Geochemists commonly quantify a mineral's affinity for, or aversion to, a given element in terms of mineral/melt partition coefficients. A partition coefficient is simply the concentration of an element in a mineral divided by the concentration of the element in the silicate liquid from which the mineral crystallized. If the partition coefficients for two different elements in a mineral are the same, then that mineral has no tendency to exclude or include one of the elements relative to the other, and the ratio of the two elements should be the same in the mineral as in the magma from which it crystallized. Partition coefficients can be used in mathematical "models" or calculations to simulate crystallization of the magma ocean or melting of mantle rocks. Using these models, the hypotheses of magma-ocean crystallization and remelting of the lunar mantle to form mare-basalt lavas can be tested.

Using pyroxene/melt partition coefficients derived from their analyses of rare-earth elements in individual pyroxene grains in mare basalts, Shearer and Papike developed a model of magma-ocean crystallization and subsequent remelting of the lunar mantle. Their model produced lava compositions having relative europium depletions similar to those of the real mare lavas, without the necessity of removing plagioclase before the olivine and pyroxene of the mantle crystallized. If this model were to survive scientific scrutiny, it would weaken one of the fundamental arguments supporting the magma-ocean hypothesis.

Orthodoxy defended: Conflicting conclusions from rare-earth elements. A potential flaw in their arguments is that Shearer and Papike could not prove that the europium depletion they measured in mare-basalt pyroxene was a result of the intrinsic ability of pyroxene to discriminate between europium and other rare earths, as opposed to merely reflecting a preexisting europium depletion in the magma from which the pyroxene crystallized.

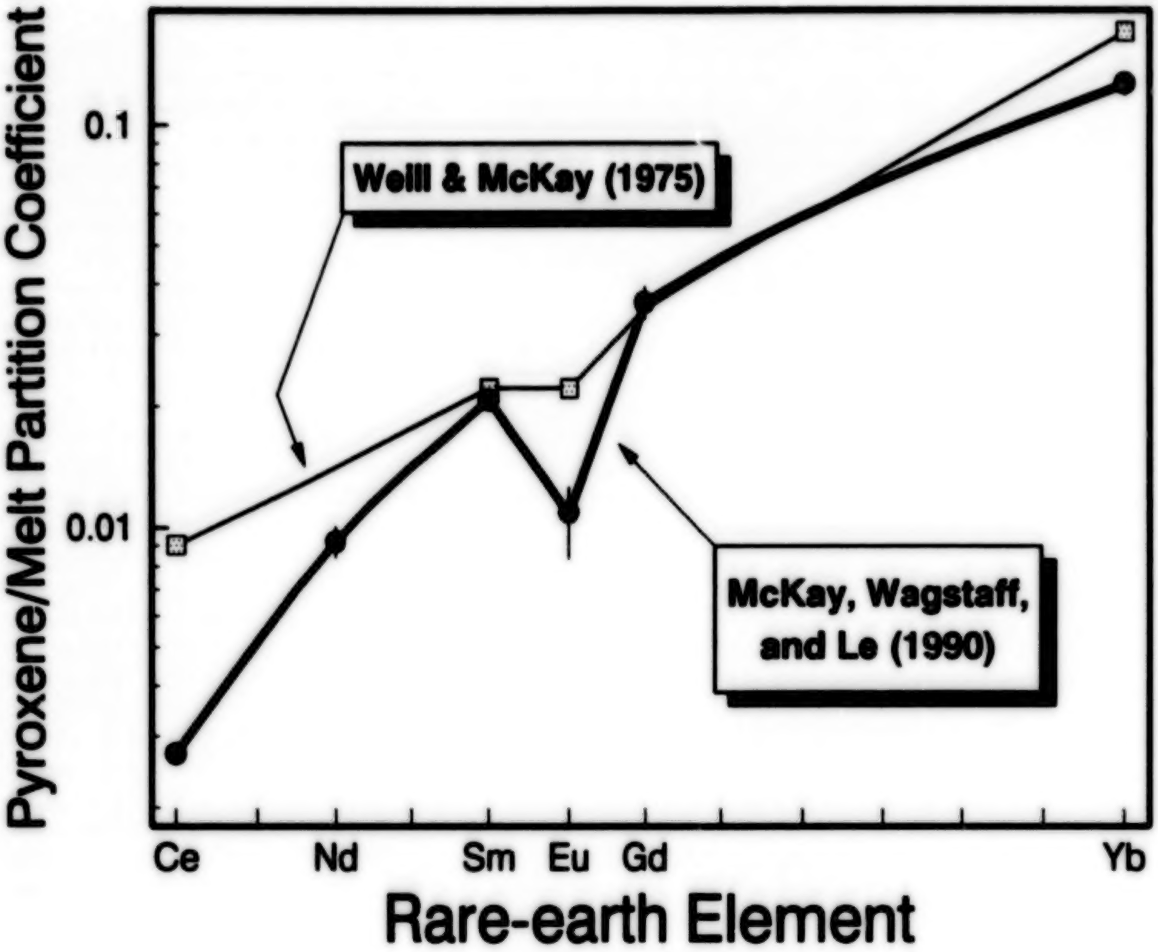

Fig. 3. Rare-earth element partition coefficient patterns for pyroxene. Partition coefficients indicate the relative tendency of elements to be incorporated in a mineral crystallizing from a silicate melt. Partition coefficients for closely related elements, such as the rare earths, are often plotted together in diagrams such as this so that any systematic behavior, or pattern, in their partitioning behavior will be evident. Both partition coefficient patterns in this diagram indicate that pyroxene has a tendency to incorporate rare earths on the right side of the diagram (those with higher atomic number) to a greater extent than those on the left, a trait long known to geochemists. The older partition coefficient pattern measured by Weill and McKay indicated little capacity for pyroxene to separate europium from the adjacent elements in the diagram. However, the newer pattern, from our recent work, shows a pronounced "dip" at europium, indicating that pyroxene has a much larger tendency to discriminate between europium and neighboring rare-earth elements than formerly believed. Implications of this tendency for the magma-ocean hypothesis are discussed in the text.

Using partition coefficients derived from other studies, James Brophy and Abhijit Basu (Indiana University) produced a mathematical model similar to that of Shearer and Papike, except that the calculated basalt compositions had much smaller europium depletions than actually observed in mare basalts. Hence, Brophy and Basu concluded that removal of plagioclase prior to the crystallization of the olivine and pyroxene in the lunar mantle was in fact necessary to match the mare-basalt europium depletion, thus supporting the magma-ocean hypothesis.

Europium partition coefficients revisited: Is the question settled? Hence the controversy hinged on exactly how strongly pyroxene can exclude europium relative to the other rare earths.

In most studies that have addressed this question, workers have relied on decade-old values for pyroxene partition coefficients, particularly values measured in rock-melting experiments by Dan Weill and myself (then at University of Oregon). These measurements indicated little tendency for pyroxene to exclude europium preferentially. However, experimental and analytical techniques have improved considerably in recent years, raising the possibility that new measurements might resolve this controversy. Therefore, Jerry Wagstaff, Loan Le (Lockheed ESCO, Houston) and I recently remeasured pyroxene/melt partition coefficients. Our results (Fig. 3) showed that, as suggested by Shearer and Papike, pyroxene does indeed have a significant capacity to exclude europium relative to the other rare earths.

The origin of the europium anomaly in mare basalts is still unsettled, however. Using our new pyroxene partition coefficients, Elizabeth Shaffer, James Brophy, and Abhijit Basu (Indiana University) recomputed their earlier models. If they allowed no plagioclase removal prior to crystallization of olivine and pyroxene, they could indeed match the europium depletion observed in the natural basalts, but only by starting with unreasonably high overall rare-earth abundances in the magma ocean. If there was prior plagioclase removal, their calculations were able to produce rare-earth abundances similar to those observed in mare basalts using a reasonable assumed starting composition for the magma ocean. The only way they could match the rare-earth element abundances in natural basalts with no prior plagioclase removal and reasonable magma-ocean initial rare-earth abundances was to use pyroxene partition coefficients much different than the ones we measured.

Could the actual partition coefficient values in the lunar mantle be similar to those required by Shaffer, Brophy, and Basu's model? Possibly. Our experiments were done at low pressure, whereas the lunar mantle minerals crystallized and remelted at high pressure, deep within the interior of the Moon. The effect of pressure on partition coefficients is not yet known, but must be studied before the final chapter in this story can be written. We are beginning such a study. In the meantime, it appears that the magma-ocean hypothesis, including the requirement for crystallization of plagioclase prior to crystallization of the minerals of the lunar mantle, is still intact.

TRITON'S POST CAPTURE THERMAL HISTORY

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Was Triton captured from solar orbit by Neptune? How was this accomplished? And what evidence exists from Triton's composition and geological evolution to support the capture hypothesis? These are fascinating questions, and ones that bear directly on the larger issues of outer solar system composition, formation, and evolution.

Planetary scientists have long suspected that Triton was captured because of its anomalous retrograde (backwards) orbit. Arguments have also been made that Triton and the planet Pluto are similar bodies. Thus, when Triton's rock/ice ratio, based on the density determined by Voyager, turned out to be very similar to that of slightly smaller Pluto, this was taken as further evidence that Triton was captured. The density of Triton was in fact successfully predicted using the Pluto analogy.

Triton's retrograde orbit, tilted with respect to Neptune's equator, can be explained if Triton is a captured satellite. Two ways for this capture to occur have been suggested. The first involves gas drag, whereby a solar-orbiting Triton passed close to Neptune when the planet was forming and thus when Neptune had a large amount of gas (mainly hydrogen and helium) in close proximity. Gas drag would slow Triton down just enough that it would become permanently gravitationally bound to Neptune. Orbital evolution would continue after the initial capture, and in principle, Triton's orbit could have shrunk and completely circularized, attaining close to its present orbit by gas drag alone.

However, unless gas drag halted, Triton's orbit would have continued to decay, so the gas drag scenario relies on the gas near Neptune having been dispersed before this could have happened, leaving Triton stranded in a retrograde, but uncircularized orbit. The gas could have been dispersed by the T Tauri wind that is thought to have cleared away the solar nebula. Thus, Triton's capture by gas drag is most plausible if it occurred close to the end of Neptune's formation, when accretion is largely over, but in which there remains a compact disk or nebula of gas and solids around Neptune. Such a nebula might, under ordinary circumstances, have given birth to a satellite system similar to that seen today around Uranus, but the capture process apparently prevented satellite formation or destroyed preexisting satellites, except close to Neptune.

The other, and perhaps more likely, capture mechanism (proposed by Peter Goldreich and co-workers at Caltech) is collision with an original regular satellite, one that would have formed from the Neptune nebula just described. Such a collision would have caused Triton to move from solar orbiting to Neptune orbiting in a single cataclysmic step. The post-collision orbit would have been a very elongated ellipse, taking Triton from several Neptune radii from the planet to greater than 2000 Neptune radii.

The post-gas-drag orbit would have been similar, but probably not so elongated. After the collision, or after the gas is dispersed in the gas drag scenario, orbital evolution would have been mainly controlled by tides, although further collisions may have occurred. In 1984, I pointed out that a captured Triton would likely have undergone massive heating as its orbit was circularized and shrunk by the tides raised on it by Neptune. I estimated that Triton completely melted, with the most volatile components of its possible makeup (methane, nitrogen) driven to the surface. These volatile ices had been identified spectroscopically at the time, and their presence on Triton's surface has been confirmed by Voyager.

The time scale of evolution is an important quantity to estimate. The simplest model of the dissipation of tidal energy assumes that Triton remains solid throughout its orbital evolution. It predicts that a Triton with an extremely elongated elliptical orbit, extending to the edge of Neptune's gravitational sphere of influence, would have taken almost 10^9 years or longer to evolve inward and have its orbit circularize. This time scale is much longer than the estimated duration of heavy cratering in the outer solar system (~500

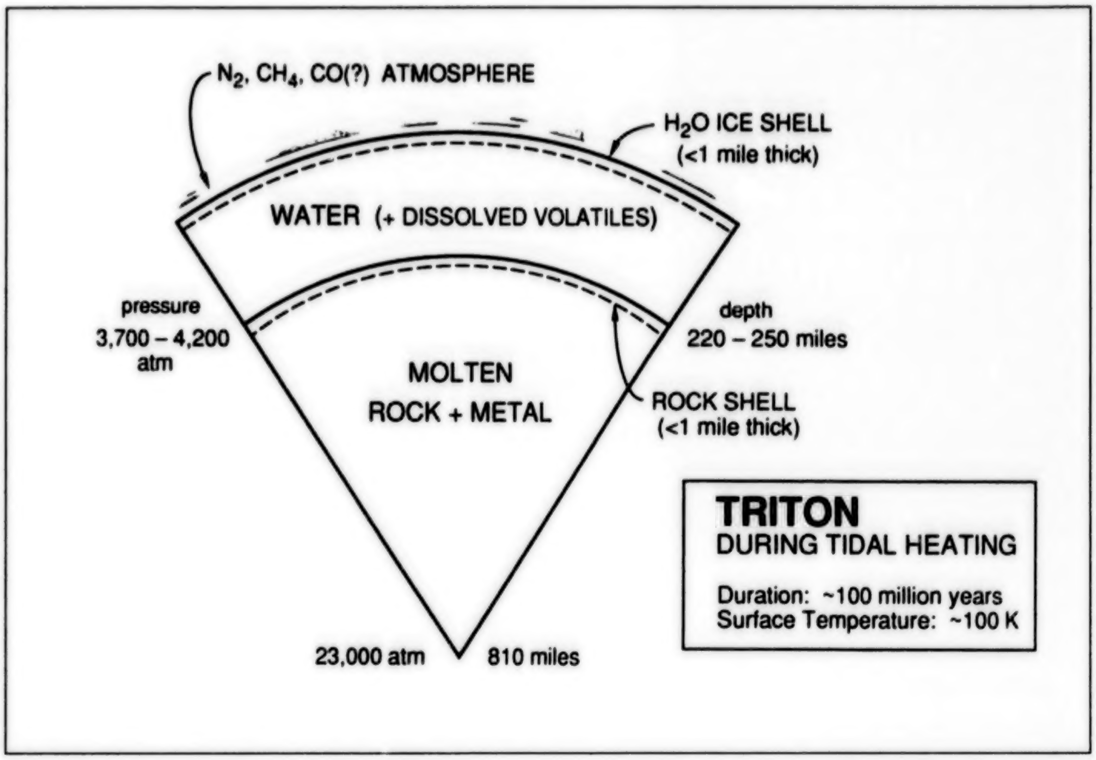
N_2, CH_4, CO(?) ATMOSPHERE
H_2O ICE SHELL (<1 mile thick)
WATER (+ DISSOLVED VOLATILES)
pressure 3,700 – 4,200 atm
depth 220 – 250 miles
MOLTEN ROCK + METAL
ROCK SHELL (<1 mile thick)
TRITON
DURING TIDAL HEATING
Duration: ~100 million years
Surface Temperature: ~100 K
23,000 atm
810 miles

Fig. 1. Schematic internal structure of Triton during the era of extreme tidal heating. For simplicity, only water ice and rock layers are explicitly noted.

million years). Thus, Triton would have stayed hot well beyond the era of heavy cratering, and it could be predicted that little if any of the heavily cratered terrains seen on nearly all of the other icy satellites would be observed on Triton by Voyager. Models that assume Triton remained solid cannot be correct, however, because so much of Triton's orbital energy would have been dumped into the satellite that it must have melted. More complex models are necessary.

How soon did Triton begin to melt once tidal evolution started? Triton is so rock-rich, approximately 70% rock by mass with the other 30% being ices, that it is likely that Triton would have begun melting its ices and unmixing, or differentiating, spontaneously, because of the energy liberated within the satellite by radioactive element (uranium, potassium, and thorium) decay. Even for very elongated initial orbits, the time-average of tidal energy dissipated would have been comparable to the power due to radioactive element decay. Therefore, it is very unlikely that melting could have been significantly delayed. Only if Triton was captured cold would melting have been put off, perhaps ~100 million years. The best estimate is that Triton would have begun to melt promptly.

Melting and differentiation inside Triton would have been partially self-sustaining because of the gravitational energy liberated by moving rock to the center and ice to the outside. However, continued radiogenic and tidal heating would have been necessary to push Triton all the way to complete differentiation. This would have been greatly aided by the fact that the tidal dissipation in a differentiating body increases dramatically compared with that in a solid one. The dissipation increases because of abundant "hot" ice (ice near the melting point) and transient multiphase regions such as ascending slush and descending mud diapirs (all of which are inherently very dissipative, and because the opening up of an internal ocean allows greater tidal flexing. Calculations suggest that Triton would have differentiated in under 10^7 years, and possibly much less. The result would have been a liquid water ocean, approximately 200-250 miles (350-400 km) deep, capped by a thin, conductive ice shell, overlying a rock core.

Continued tidal heating in the core would have caused it to heat up and melt as well. The ultimate tidally heated configuration for Triton would have been nearly totally molten (Fig. 1). A thin water-ice shell would have topped a liquid water mantle, and a thin rock shell would have topped a molten silicate core; this lower shell may have been negatively buoyant, though, and could have turned over as on a lava lake. There may have been an inner core of liquid iron-sulfur, but no iron shell because the freezing point of the core would have been less than that of the molten rock mantle. There might have been substantial ammonia (NH_3) dissolved in the water mantle, and other ices such as methane (CH_4) and nitrogen (N_2) might have formed a surface ocean.

Once Triton melted, and it needed only a small portion of the total orbital energy potentially available to it, its orbital evolution would have slowed. This seeming paradox results from the fact that dissipation in a molten Triton would have been confined to the thin solid shells just described. The hot, near-melting points of both shells would have been quite dissipative under tidal forcing, but the shells would have been at most only a few miles thick. Tidal flexing is at its maximum for a liquid body, but the total effect would have been to stretch out Triton's orbital and thermal evolution due to tidal heating. Thus a nearly molten Triton may have stayed hot for an extended length of time, greater than 500 million years. Therefore, as with the simple model of tidal heating above, it is predicted that Triton should not have retained any early record of heavy cratering.

The real question is whether tidal heating and melting are required to explain the absence of heavily cratered terrains on Triton. This depends on the composition of the surface. If at least portions of Triton's surface are made of nearly pure water ice, then ancient craters could survive in these cold, rigid regions in the absence of tidal heating and associated geologic activity. Thus tidal heating would be necessary to explain Triton's present appearance. If, however, there is a crust of lower melting point ices (such as nitrogen, methane, or ammonia) greater than a few miles thick, then continuing volcanic and other activity in this relatively soft layer driven by present-day radiogenic heating inside Triton could probably destroy any ancient crater population. Similarly, a liquid layer may be maintained in the mantle by the presence of ammonia. In these cases, tidal heating may not be required.

In summary, the results of recent calculations and modeling suggest that massive tidal heating and melting, associated with Triton's capture by Neptune, could be largely responsible for Triton's geologically youthful appearance. However, in order to determine whether massive tidal heating actually occurred will require detailed geological analyses to constrain the composition of Triton's surface layers. Further evidence for or against the capture hypothesis will come from studying the chemical evolution of Triton, to see if the detailed chemical composition of this satellite is consistent with an era of extreme tidal heating.

BROILED ALIVE! AN INCENDIARY THEORY OF THE DINOSAURS' DEMISE

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A solitary meteor streaks across the dark night sky. The momentary surprise and pleasure of glimpsing this display is shared by people over an area a few hundred miles square. The source of the light is usually a tiny fragment of extraterrestrial matter less than one half an inch in diameter plummeting into the Earth's atmosphere at a speed exceeding 25,000 miles per hour. The friction between this fast-moving fragment and the Earth's air raises its temperature above the 3600° F vaporization point of metal or rock, and the incandescent meteor begins to boil away. Most meteors evaporate completely before they reach a height of 40 miles, although rare large meteors may slow down enough so that some of their material can reach the ground. We do not feel the intense heat produced by an individual meteor because it is so far away.

Now, instead of the beauty of a single meteor, imagine a sky full of meteors. During the so-called meteor showers hundreds or thousands of meteors may appear each minute and at any given time several meteors may be visible simultaneously. Such displays are nowadays regarded as rare and beautiful, but not particularly dangerous, natural phenomena like rainbows or auroras. But now imagine a meteor shower so intense that the individual meteors cannot be resolved by the eye and the entire sky blazes with the dazzling light of hundreds of millions of tiny rock fragments, each one heated to vaporization. The aggregate heat of all these meteors is unbearable to observers on the ground. Steam begins to hiss from overheated leaves while dead twigs and plant stems burst into flame. The heat continues to build. Green trees now burst into flame and the air fills with choking smoke. Above it all the sky continues to glow a relentless blue-white while nearly everything below is burnt into ash.

Is this a scene from a science fiction story? Maybe not, according to recent investigations of the effects of large impacts on the Earth. This may, in fact, be what the dinosaurs saw on their last day of existence 65 million years ago at the end of the Cretaceous Era, the last Age of the Dinosaurs.

The modern story of the death of the dinosaurs began in 1980 with the publication of a paper by University of California at Berkeley physicist Louis Alvarez and geophysicist Walter Alvarez, father and son, along with chemists Frank Asario and Helen Michel (Lawrence Berkeley Laboratory). These scientists had identified a 1/8" to 1/2" thick layer separating Cretaceous rocks from Tertiary rocks at Gubbio, Italy. In this layer they found a strong enrichment of the rare element iridium. Although iridium is rare in the Earth's crust, it is much more common in meteorites, and the Berkeley group concluded that the layer had been laid down in the aftermath of the impact of a 6-mile diameter asteroid or comet somewhere on the Earth.

In subsequent years, the iridium-rich layer has been found at more than 100 other sites distributed over the globe, in both marine and land environments, and it seems to be uniquely associated with the end of the Cretaceous: no other such widely distributed iridium-rich layers have been found anywhere else in

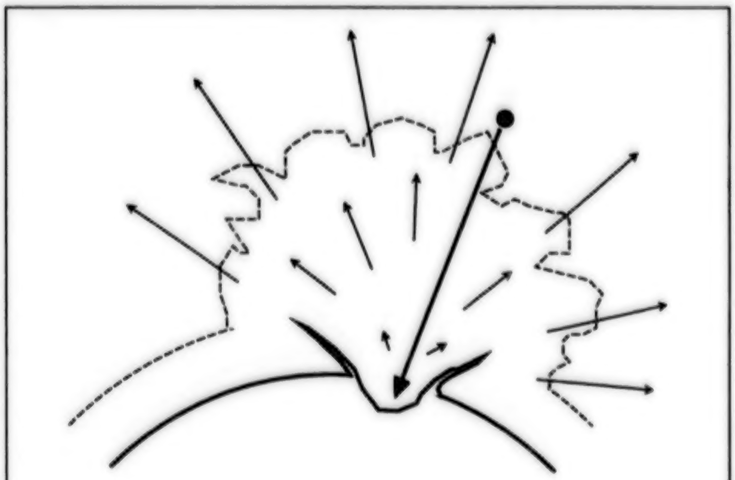

Fig. 1. The impact of an asteroid or comet creates a plume of hot vapor and melt that expands away from the impact site at high speed. If the impact is large enough the vapor plume blasts the atmosphere aside and the hot debris vents above the atmosphere.

the geologic record. Other investigations have refined our knowledge of the layer and its contents. Although it is altered at nearly all sites, Jan Smit and G. Klaver (Geological Institute, Amsterdam, The Netherlands) in 1981 showed that at the best-preserved locales the boundary layer is full of tiny spherules, each about the size of a grain of sand. Wendy Wolbach and Ed Anders (University of Chicago) in 1985 discovered soot and charcoal fragments in the layer in such great abundances that all of the Cretaceous forests must have burned in a single, global wildfire about the same time the iridium-bearing spherules were deposited. Evidence of increased strontium deposition in the oceans suggests that intense acid rain may have washed the continents and upper oceans.

Although the actual crater caused by the impact has not yet been found (recent searches are centering on the Caribbean), the evidence for a large impact at the end of the Cretaceous has become nearly overwhelming: besides the iridium, other characteristic meteoritic elements are found in the boundary layer. Tiny fragments of shocked quartz and the high pressure minerals coesite and stishovite have been detected. Stishovite, in particular, can only be produced by the immense pressure of meteorite impact, and has long been accepted as the ultimate criterion for impact.

Even if we accept the physical evidence for a large impact at the end of the Cretaceous, a link is still missing in the story of the dinosaurs' death: how did they (and the myriads of other species that became extinct at the same time) actually die? In their original paper, the Alvarezes proposed that the dust lofted by the impact might fill the lower atmosphere, darkening the skies and bringing photosynthesis to a halt. This dust, they suggested, would be globally distributed by the winds and the entire Earth would be plunged into darkness for periods of months. The giant land animals, victims of disrupted food chains, would die a slow death by starvation. The problem with this "darkness at noon" scenario (which led directly to the "nuclear winter" view of the effects of global nuclear war) is that the impact would have to produce an enormous quantity of very fine (less than 0.00004 inch diameter) dust.

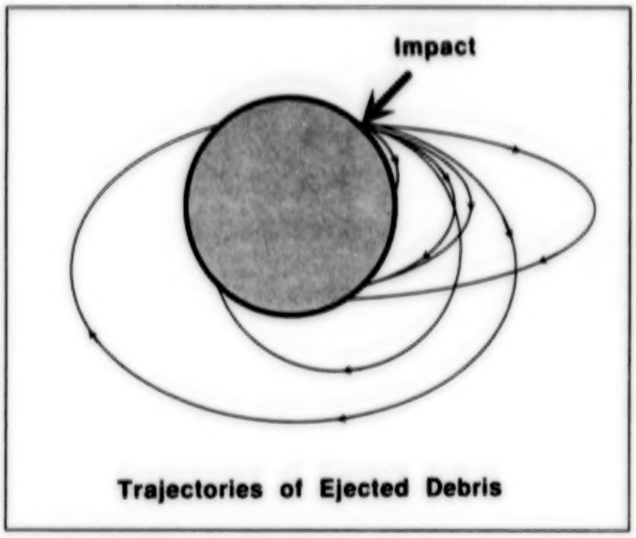

Fig. 2. Condensed debris from the vapor plume follows ballistic trajectories that re-impact the Earth at great distances from the initial impact site. Because debris may travel different paths to get to the same site, it arrives gradually over a period of about an hour after the impact.

Unfortunately, there is no evidence, either experimental or theoretical, that the necessary quantities of fine dust could be made by an impact. Even worse, the boundary layer itself seems to be largely composed of spherules the size of sand grains—much too large to be blown significant distances by the wind. Sand-size particles fall out of the atmosphere in a few hours, making it very unlikely that they could have been suspended for periods of months. The darkness at noon scenario also does not explain the evidence for global wildfires.

A new view of the events at the end of the Cretaceous has recently grown out of studies of the effects of a large impact on the Earth. It is now recognized that the plume of hot gas and melted projectile created by a high-speed impact would not have stopped at mere stratospheric heights. Instead, the violently expanding vapor plume from a 6-mile diameter asteroid or comet would have blasted the atmosphere aside, spewing melt and vapor into space well above the atmosphere (Fig. 1). The vapor, cooling as it expands, condenses into myriads of tiny droplets that follow ballistic-missile-like paths from the impact site to all parts of the Earth (Fig. 2).

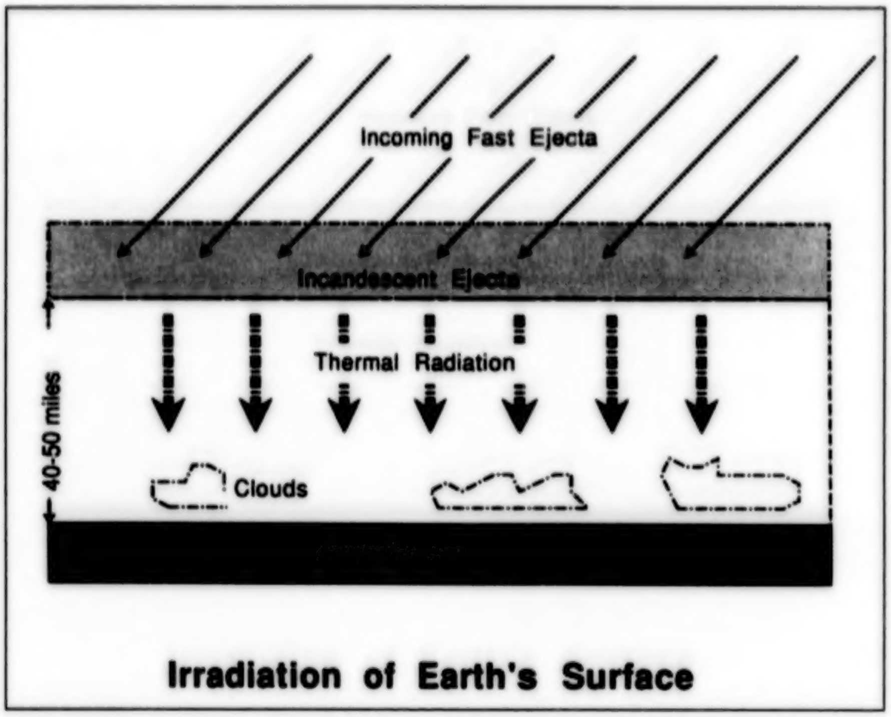

Fig. 3. When the fast ejecta arrives at the top of the atmosphere, friction with the air slows it at altitudes of about 45 miles, turning its kinetic energy into heat. The radiant heat from the ejecta may be intense enough to ignite forest fires on the ground.

Theoretical studies indicate that the particles would have each been about the size of fine sand. The reentering droplets lose most of their energy at altitudes of about 45 miles, radiating it as heat (Fig. 3). Unlike normal meteors, the small, relatively slow-moving droplets do not completely vaporize, but instead drop quietly to the ground once their initial velocity is spent.

Computations by myself, Nick Schneider (University of Colorado), Kevin Zahnle (NASA Ames), and Don Latham (U. S. Forest Service) show that the amount of radiant heat reaching the ground from a Cretaceous-size event is about 10 times the heat of the Sun at noon on a hot day in Death Valley. It is nearly equal to the heat per unit area produced in an oven turned to "broil". This bath of intense heat would last more than an hour over most of the Earth's surface, peaking some twenty minutes after the first ejecta began to arrive. Even green trees on the ground would burst into flame after about half an hour of this broiling. Normal clouds could provide little protection: only the thickest clouds contain enough water to prevent the radiation from burning through and searing the ground below.

In addition to the radiant heat, the reentering ejecta heats the upper air to high temperatures, allowing the nitrogen and oxygen to combine into NO, which then mixes downward. Combining with H_20 in the lower atmosphere, massive amounts of HNO_3, nitric acid, form. Further studies by Ron Prin (MIT), Bruce Fegley (Max Planck Institute, Mainz), and Kevin Zahnle suggest that acid rain in unprecedented amounts would sweep the burned-out forests and the upper ocean water would become so acid as to cause the death of the teeming plankton populations of the Cretaceous seas.

Apocalyptic as these events seem, they are all reasonably well supported by the evidence from the Cretaceous-Tertiary boundary layer. The pulse of radiant heat is an inevitable consequence of the reentry of far-flung debris from the main impact, if it is once accepted that most of the extraterrestrial material in the boundary layer was emplaced ballistically, rather than by wind dispersal. Although charred dinosaur bones have yet to be found, the boundary layer contains abundant soot and even some small lumps of charcoal that bear testimony to the global wildfires that raged during deposition of the layer. The true wonder of these events is that anything at all survived the catastrophe. Our ancestors, the mammals, may have survived because they lived in deep burrows and could escape the heat and flames. The large-bodied dinosaurs were evidently not so lucky.

Much work still needs to be done on the effects of large impacts on the Earth and the means by which they have affected the biological world. The new field of catastrophic geology opened by the Alvarezes in 1980 is still expanding and providing new insights into how our planet, and the solar system, really works.

GEOLOGICAL TIME ON THE MOON: THE AGE OF THE CRATER AUTOLYCUS

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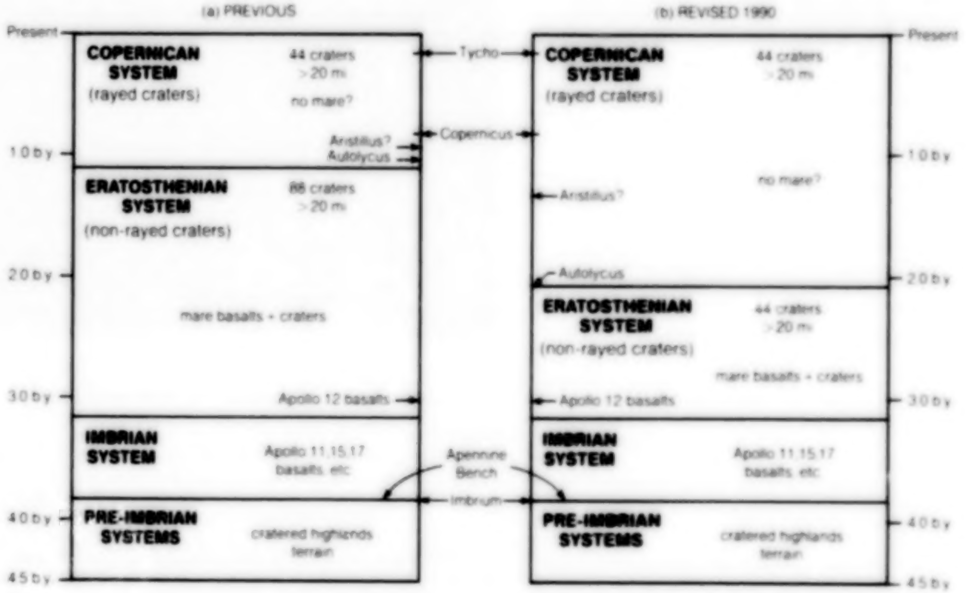
The formation of the Imbrium basin 3.85 billion years ago by a giant impact on the surface of the Moon coincided with a change in the surface evolution of our satellite. Terrains formed before the Imbrium impact are heavily cratered highlands. Terrains formed after the Imbrium impact are less intensely cratered, and they include lava flows of mare basalt that flooded low places on the lunar surface, creating broad dark plains visible from the Earth with the naked eye.

Lunar surface units formed after the Imbrium impact are assigned to three systems, that is, different time periods (Fig. 1). The divisions are relative and determined by several characters: 1) younger units overlie older units; 2) older units have more craters; and 3) younger craters are better preserved. The oldest of these systems, called the Imbrian, is the most heavily cratered and includes the flood plains of mare lava that were sampled by the Apollo 11, 15 and 17 missions. The intermediate-aged system, called the Eratosthenian, is less cratered and includes the mare lavas at the Apollo 12 landing site. The youngest system, called the Copernican, is least heavily cratered and may entirely postdate eruption of mare basalt onto the lunar surface.

The working distinction between the latter two of these systems is that Copernican craters larger than a few miles in diameter still have visible rays, whereas Eratosthenian craters have had their rays eroded away by later small cratering events. (Rays are light-colored streaks emanating from an impact crater and they consist of thin deposits of materials ejected from that crater when it formed.)

Determination of the solidification ages of mare basalts collected at the various Apollo landing sites allows us to infer that the transition between the Imbrian and Eratosthenian periods took place about 3.2 billion years ago. Unfortunately, the time of the transition between the Eratosthenian and Copernican


(a) PREVIOUS
(b) REVISED 1990
Present
1.0 b.y.
2.0 b.y.
3.0 b.y.
4.0 b.y.
4.5 b.y.
COPERNICAN SYSTEM
(rayed craters)
44 craters >20 mi
no mare?
Aristillus?
Autolycus
Tycho
Copernicus
ERATOSTHENIAN SYSTEM
(non-rayed craters)
88 craters >20 mi
mare basalts + craters
Apollo 12 basalts
IMBRIAN SYSTEM
Apollo 11,15,17 basalts, etc.
Apennine Bench
Imbrium
PRE-IMBRIAN SYSTEMS
cratered highlands terrain
COPERNICAN SYSTEM
(rayed craters)
44 craters >20 mi
no mare?
Aristillus?
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ERATOSTHENIAN SYSTEM
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44 craters >20 mi
mare basalts + craters
Apollo 12 basalts
IMBRIAN SYSTEM
Apollo 11,15,17 basalts, etc.
PRE-IMBRIAN SYSTEMS
cratered highlands terrain

Fig. 1. Division of the lunar surface units into systems, with corresponding time scales indicated. a) Commonly accepted view, after Don Wilhelms, United States Geological Survey, with the Copernican system starting about 1.1 billion years ago. b) Our revised view, taking into account our conclusion that the Copernican crater Autolycus formed 2.1 billion years ago.

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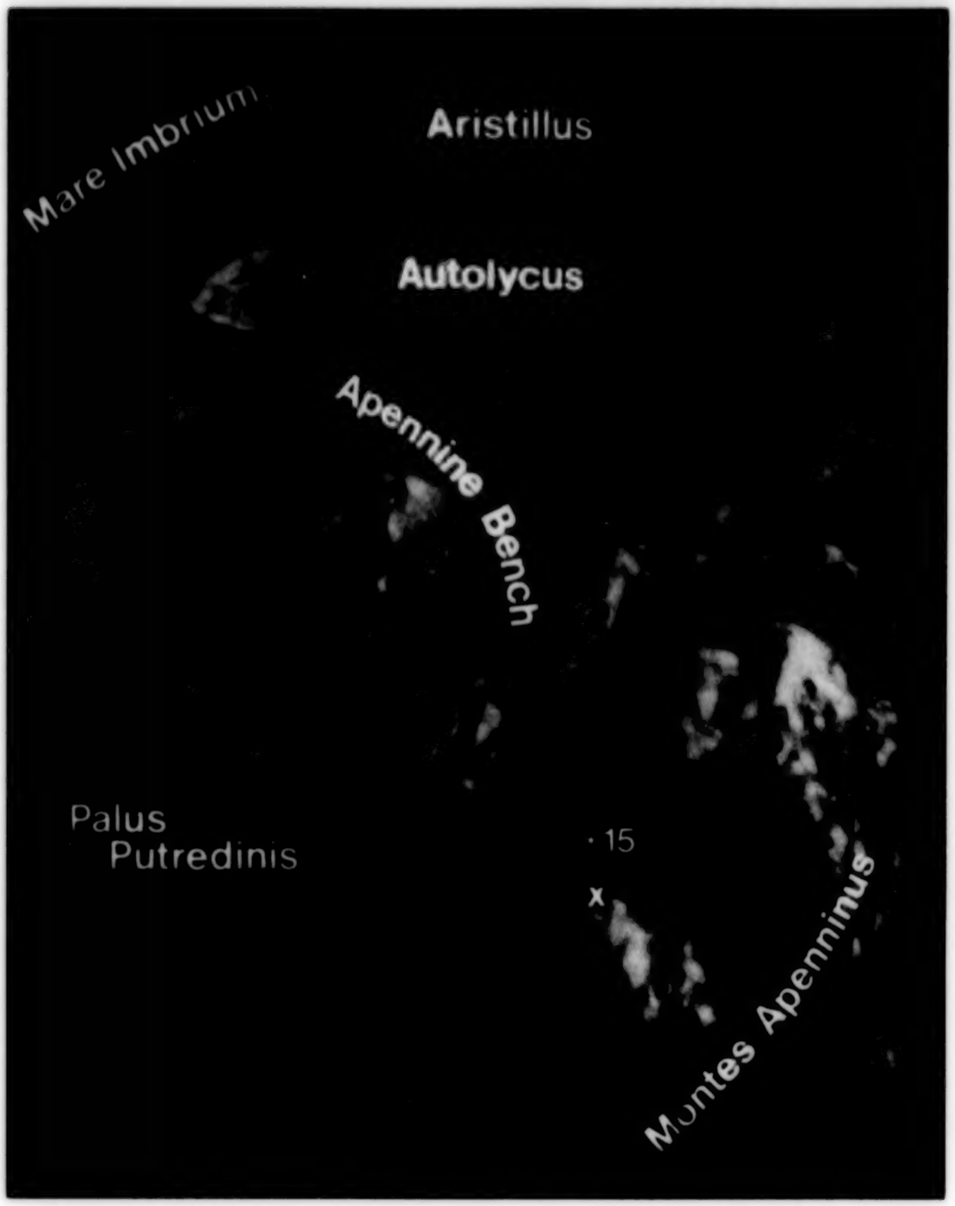

Fig. 2. The region of the Apollo 15 landing site (NASA photograph; Apollo 15 frame M-1537). "15" marks the area of the landing site; "x" marks the specific collection area for the KREEP basalt samples we studied. The direction of view is toward the north. The distance from the Apollo 15 landing site to the rim of Autolycus is about 80 miles; Autolycus is about 30 miles in diameter. The Apennine Bench is a plains unit that was formed immediately after the giant impact that formed the Imbrium basin; this unit consists of KREEP basalts.

periods is more difficult to determine, because none of the Apollo sampling sites were on units formed after this transition.

The time of the Eratosthenian-Copernican transition has been estimated by assuming that impact craters have formed on the lunar surface at a constant rate over about the past 3 billion years. The date of the transition so derived is 1.1 billion years ago. This 1.1-billion-year age is consistent with the age inferred for the crater Copernicus, which is about 800 million years.

If the transition from Eratosthenian to Copernican took place 1.1 billion years ago, the Eratosthenian period would represent about twice the time of the Copernican period, because it contains twice as many craters larger than about 20 miles across (Fig. 1a). The actual age of the Eratosthenian-Copernican transition is important to determine because it will define the length of time mare basalts were erupted onto the lunar surface, and thus will give us information on the internal thermal history of the Moon. It also will help determine whether or not the rate of cratering on the lunar surface has changed with time, thus giving us information on the flux of small bodies in the inner solar system over the last 3 billion years.

Our investigations of the thermal history of three small samples from the Apollo 15 landing site, using the Ar-Ar method of radiometric age determination, permit revision of the estimate of the time of the Eratosthenian-Copernican transition. The samples we studied were all small chips of potassium-rich lavas (called KREEP basalts by lunar petrologists). KREEP basalts are not representative of the lavas that form the mare plains at the Apollo 15 landing site. Instead, they are older, having solidified about 3.85 billion years ago, immediately after the great impact that formed the Imbrium basin. They apparently were later ejected from the area where they solidified and were thrown onto the surface at the landing site by an impact.

Our analyses demonstrate that all three samples were heated and altered by an impact that took place 2.1 billion years ago. It is likely that this was the same impact that brought the rocks to the site. The question is: What impact was it that heated, altered, and transported these fragments?

Two large craters that formed during the Copernican period, Aristillus and Autolycus, lie to the north of the Apollo 15 landing site (Fig. 2). The shape of Autolycus is degraded and Aristillus appears fresher, so that Aristillus must be younger. Even prior to the Apollo 15 mission, it was recognized that light-colored streaks, called rays, emanating from one or both of these craters crossed the landing site. These rays are thought to consist of fragments ejected from the craters and sprayed out across the lunar surface. Thus, rock fragments from Aristillus and/or Autolycus should be present at the Apollo 15 site.

Autolycus is our prime candidate for a source for our samples of altered KREEP basalt. It is in an appropriate terrain, the Apennine Bench, which has been established by remote sensing of chemical composition to consist of KREEP basalts. It is also closer to the landing site than Aristillus. Craters farther away than these two should not have transported much material to the landing site, and there are no closer craters of appropriate size to be the source of the samples we have analyzed. Thus, our sample data and the geology of the region strongly suggest that the crater Autolycus formed in an impact 2.1 billion years ago. The only plausible alternative is that Aristillus is 2.1 billion years old, in which case Autolycus would be even older.

Autolycus is one of the most degraded, thus oldest, of the craters formed during the Copernican period. If it is 2.1 billion years old, as our data strongly suggest, then the generally accepted lunar geologic time scale needs revision, as shown in Figure 1b. We conclude that the Copernican period was twice as long as the Eratosthenian, not half as long. Our results indicate that the assumption of a constant rate of impacts onto the lunar surface for the past 3 billion years is incorrect, as we find that the Eratosthenian craters, which are twice as numerous as the Copernican craters, formed in only half the time. Thus, it appears that the cratering rate was four times as high in the earlier Eratosthenian period as in the Copernican.

Our results further indicate that the flooding of the lunar surface by mare basalts may have ceased about 2 billion years ago, not 1 billion years ago. If this is the case, then the source of internal heat that produced the basaltic lavas ceased to be effective much earlier than previously thought.

VENUSIAN HIGHLANDS: THEIR GRAVITY AND TOPOGRAPHY

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As the planet most similar in size and density to Earth, Venus provides an invaluable testing ground for our understanding of dynamic processes on Earth. One of the fundamental questions for Venus is how the planet loses its internal heat. Earth loses its heat primarily through the system of plate tectonics. Hot material rises to the surface at oceanic spreading ridges; cold material returns to the interior at subduction zone trenches. The pattern of rising and sinking material is related to convection in the interior. Since silicate bodies significantly smaller than Earth (like Mars, Mercury, and our Moon) contain a smaller volume of heat producing elements, they are able to lose their heat primarily through conduction to the surface. Given the similarity in size and mean density between Venus and Earth, we expect that Venus may lose most of its internal heat in the same manner as Earth, namely through plate tectonics.

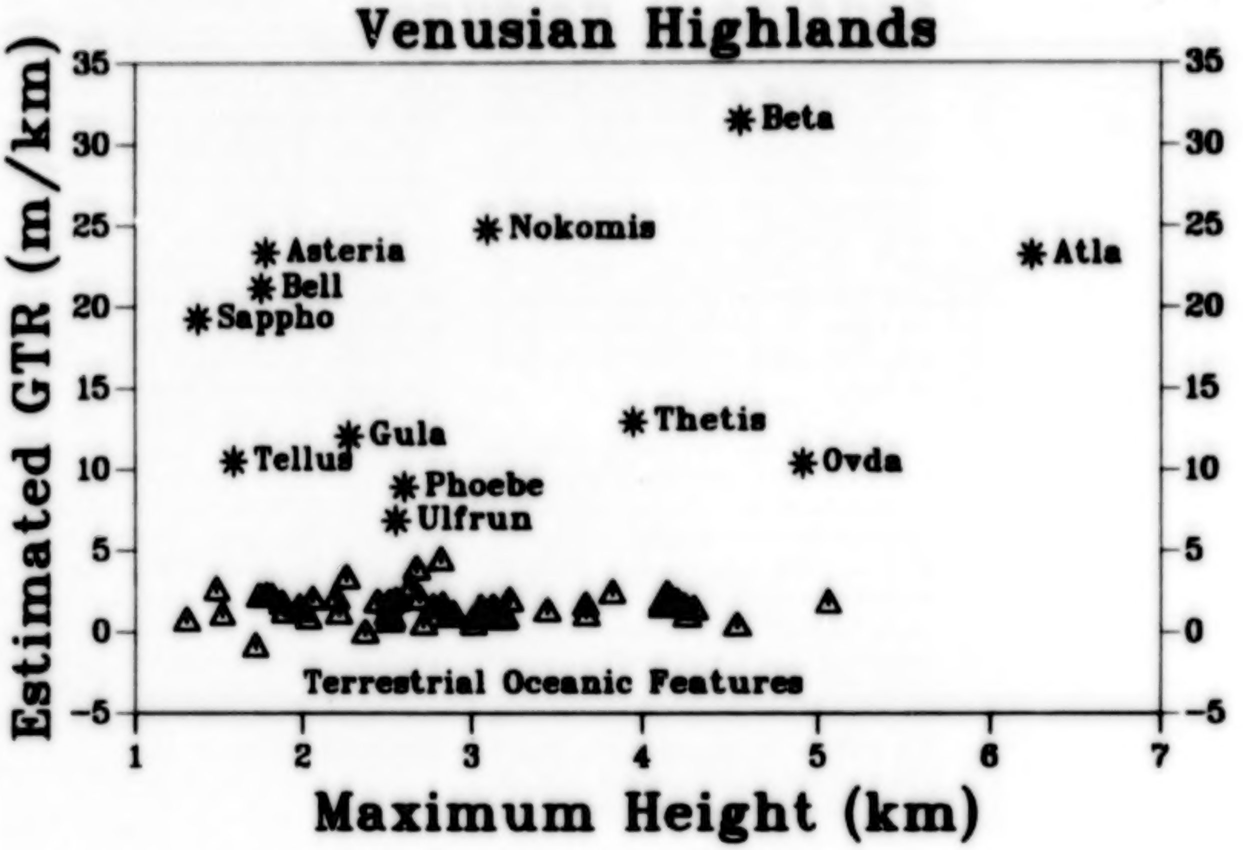
The most obvious surface manifestation of plate tectonics is an extensive network of trenches and spreading ridges. Although currently available data for Venus does not rule out the existence of ridges and trenches, it seems clear that if plate tectonics operates on Venus, it does so in a less vigorous manner than on Earth. Earth also loses some of its heat at "hot spots" such as Hawaii. Hot spots consist of regions where hot material rises in an isolated convective plume, rather than at a linear ridge, causing enhanced heat flow and volcanism. One hypothesis is that on Venus, in contrast to Earth, hot spots are the primary heat loss mechanism, and plate tectonics plays a secondary role. The morphology of many venusian highland regions is similar to terrestrial hot spots, consisting of broad, gentle domes, with smaller, steeper peaks that appear to be volcanic.

Large topographic masses, such as those found at hot spots, exert stress on the interior of a planet. Often this topographic stress is balanced by material that is lower in density than the surroundings, and which gives rise to forces directed upwards. Many terrestrial mountains are supported by low density "roots" that extend down into the more dense material normally found in the interior. This mechanism is known as Airy compensation.

The topography above hot spots is supported, or compensated, by both the dynamic buoyancy force of the rising plume ("dynamic compensation") and the density contrast that occurs when the plume heats up cooler material near the surface ("thermal compensation"). The depth at which the pressure of the excess mass of the mountains is balanced out is termed the "compensation depth." By modeling the relationship between gravitational acceleration and topography, it is sometimes possible to distinguish between compensation mechanisms. For instance, gravity data for Venus suggest that the topography at several highland regions is compensated dynamically by convective flows, lending strength to the idea that they may be hot spots.

In a recent study we used gravity data to estimate compensation depths for 12 venusian highland features: Asteria, Atla, Bell, Beta, Ovda, Phoebe, Tellus, Thetis, and Ulfrun Regiones, Nokomis Montes, Sappho and Gula Mons. The compensation depth is found by comparing the observed and predicted gravity, which consists of the gravity expected from the topography plus the contribution from the compensating mass at depth. The model of predicted gravity is rerun until a best fitting compensation depth is found. Compensation depths for the 12 highland regions range from 50 km (31 miles) to 270 km (167 miles).

In addition to compensation depth, the geoid to topography ratio provides useful information for distinguishing between compensation mechanisms. The geoid is the integral of the gravity data, and is difficult to calculate directly from the available data for Venus. Using the compensation depth and the topography, the geoid for each region can be estimated. The geoid to topography ratio (GTR) range for the venusian features named above is 7-31 m/km (37-164 ft/mile). The GTR values fall into two



statistically distinct groups: a lower GTR group with a range of 7-13 m/km (37-69 ft/mile), and a higher GTR group with a range of 19-25 m/km (101-132 ft/mile). The GTRs for the 12 venusian highland features are plotted versus maximum topographic height in Figure 1.

Fitting theoretical models to the two GTR groups shows that the lower group can best be explained by thermal compensation. Some contribution to the support of the topography from either Airy or dynamic compensation cannot be ruled out. The upper GTR group requires a component of dynamic support, although a small contribution from Airy or thermal compensation is likely. This evidence for dynamic compensation of the highland features, along with their morphology, suggests that the regions in the upper GTR group are active hot spots. The features in the lower GTR group could also be related to hot spots, but because the proportion of dynamic support is much less, they would have to be in a different stage of evolution, perhaps dying out.

On Earth, the GTRs for topographic features supported by a variety of compensation mechanisms all fall into the range 1-5 m/km (5-26 ft/mile). Figure 1 shows GTRs for 53 terrestrial oceanic hot spots, swells and plateaus (Sandwell, University Texas, Austin). GTRs for venusian features are dramatically higher than GTRs for equivalent terrestrial regions. If the oceans were removed from those features, their GTRs would increase by a factor of ~1.5, as the density contrast at the surface would increase from the difference between crustal rock density and water to the difference between crustal rock and air.

The most significant contribution to high venusian GTRs appears to be the lack of a low viscosity zone. On Earth, between roughly 100 km (62 miles) and

Fig. 1. Estimated geoid to topography ratios as a function of topographic height. Stars indicate venusian highland features; triangles represent terrestrial oceanic hot spots, swells and plateaus.

200 km (124 miles) depth, a zone of low viscosity occurs. This zone exists because the pressure-temperature dependence of various mantle minerals causes them to be more readily deformed at those depths. Lower melting temperatures, perhaps due to the presence of water in this region, may also contribute to decreasing the viscosity. As the surface temperature on Venus is 450° C (842° F) hotter than Earth, the temperature-pressure conditions that exist in the upper few hundreds of kilometers (miles) are considerably different from those on Earth.

Thus, a low viscosity zone may not exist at all. Upper mantle convection models show that the presence of a significant low viscosity zone decreases the GTR by a least a factor of 2 (Elizabeth Robinson, Stanford). The extremely high GTRs found for some venusian highlands cannot be fit if a low viscosity zone is included in convective models. On Earth, the presence of a low viscosity zone means that GTRs for oceanic topographic features which are dynamically compensated can appear arbitrarily small; this results in a small overall GTR range.

The lack of a low viscosity zone has important implications for Venus. The low viscosity zone on Earth may play a key role in plate tectonics. Without the lubrication provided by a weak zone, the resistance to plate motion might be sufficient to stop or at least slow down plate velocities. This may mean that if plate tectonics exists, it could not be as vigorous a heat loss mechanism on Venus as on Earth. Further, without a low viscosity zone, the horizontal forces of convection might be coupled into the upper surface layers, causing the layers to deform as they resist the convective motion (Phillips, SMU). On Earth, the plates move easily over the low viscosity zone until added resistance is encountered (for example, at subduction trenches). Thus despite the similarities in bulk properties between Venus and Earth, the variation in surface temperature appears to result in important differences. The lack of a low viscosity zone appears to control which heat loss mechanism dominates vigorously convecting planets: plate tectonics or hot spots.

CRUSTAL STRUCTURE AND EARLY BOMBARDMENT HISTORY OF THE MOON

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Understanding the early evolution of the Moon is a complicated task that requires synthesis of many different kinds of information. For many research topics, this synthesis is best achieved by multidisciplinary studies involving several workers with varied fields of expertise. I am collaborating with Graham Ryder (Lunar and Planetary Institute, Houston), Jeff Taylor and Klaus Keil (University of Hawaii), and Richard Grieve (Geological Survey of Canada) in a multidisciplinary study that addresses several important topics in early lunar crustal evolution.

One of the topics we are studying is the composition and structure of the Moon's crust. Geophysical experiments carried out by the Apollo astronauts indicated that this outer layer of the Moon is composed of relatively low-density rock and is about 70 kilometers thick (43 miles). Before we can understand how and when this crust formed, we need to know the elemental composition of its rocks and the spatial distribution of different rock types.

A second topic we are investigating is the history of impact bombardment of the lunar surface. Impact, by bodies of a great range of sizes, has been one of the major processes that has shaped the lunar surface. If we can determine the times that different impact craters formed, we can construct a time scale for events in lunar surface evolution.

Our studies of the largest of the lunar impact craters have given us information on both these topics. The largest lunar craters are called basins. These are circular structures, consisting of alternating concentric rings of troughs and ridges. Their diameters range from 300 to 2500 kilometers (185-1550 miles). The impacts that formed these basins excavated rock from great depths in the crust, at least several tens of kilometers, and brought this rock to the surface. By studying the material excavated by basin impacts, we can determine the nature of the crust at depth. Furthermore, if we can determine the time when an individual basin formed, this date gives us information on the timing of other geologic events on the Moon.

Our studies have focused on investigation of samples from the lunar highlands, the light-colored, densely cratered, relatively old areas of the lunar surface. The material at the lunar surface is a kind of soil, called regolith, that consists of fragments of rock broken, mixed, and melted by small impacts. All the lunar samples brought to Earth were collected from regolith, so none were found in the places in which they formed. Some of the lunar samples collected from highlands regolith are igneous rocks — they represent the solidification products of lavas or silicate melts that formed as a result of internally generated heating of the Moon. Among these are rocks formed during the very earliest history of the Moon, and such rocks (called pristine) appear to be the primary building blocks of the lunar crust.

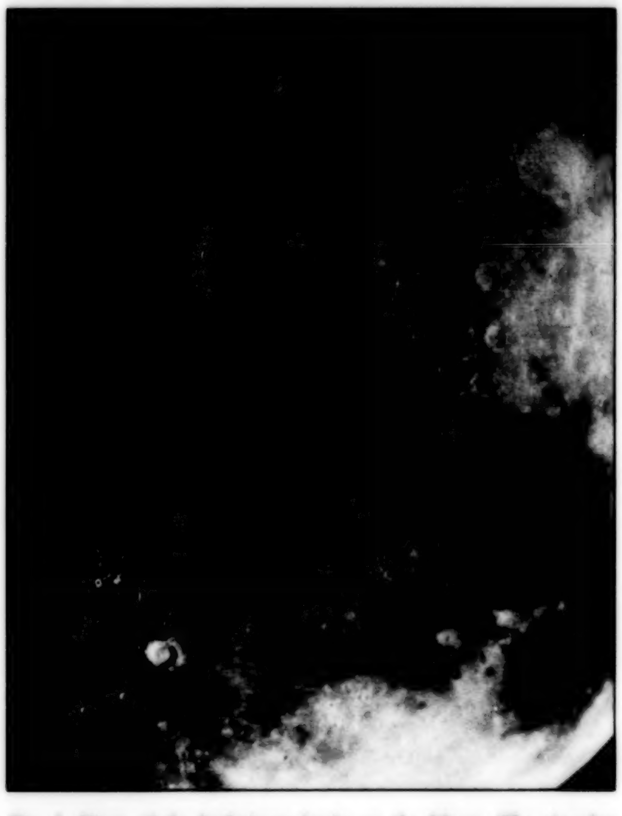

Fig. 1. View of the Imbrium basin on the Moon. The circular arrangement of mountains, scarps, and ridges delineates the rim of a giant ancient crater, called a basin, that is almost 1200 kilometers in diameter (750 miles). After the basin formed, its low-lying regions were flooded by dark lavas.

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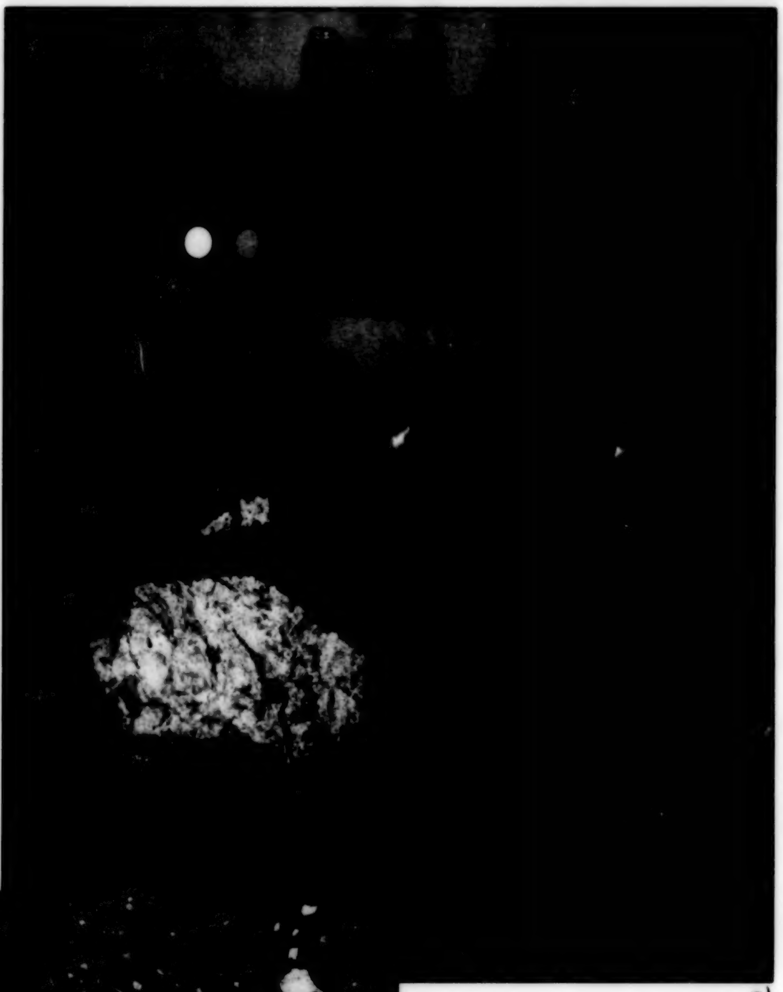

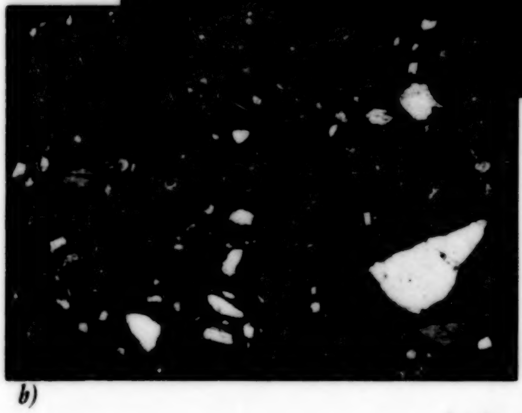

Fig. 2. Lunar sample 15455, collected by astronauts Dave Scott and Jim Irwin during the Apollo 15 mission to the Apennine mountains in 1971. a) The sample before it was subdivided for study (NASA photograph S-71-46527). The dark material is impact melt rock formed during the Imbrium basin impact. The white materials are fragments of pristine igneous rocks. b) A view of a thin slice of the dark impact melt rock, as seen through a microscope with crossed polarizers. Width of the field of view is about 2 millimeters (about 8/100 of an inch). The smaller grains represent crystallized impact melt. The larger crystals are unmelted mineral fragments that were included within the melt. We have determined the elemental compositions of these mineral fragments to determine the nature of their parent rocks.

Most of the lunar samples collected at highlands sites, however, are rocks that have been strongly altered by impact processes. Most highlands rocks are breccias — they consist of aggregates of rock fragments that have been shattered, in some cases mixed, and in some cases melted by impact processes. Others are impact melt rocks — they represent solidified silicate melt that was produced by the tremendous heating associated with impact. The origin of some of these breccias and impact melt rocks has not been fully explained, and we believe that they contain clues to early lunar evolution.

Two rocks collected from the Apennine mountains of the Moon in 1971 by the Apollo 15 astronauts appear to be breccias created by the giant impact that formed one of the largest basins on the Moon, the Imbrium basin (Fig. 1). The Imbrium basin is over 1100 kilometers (680 miles) across and its formation probably excavated rock from depths as great as 70 to 80 kilometers (43-50 miles), at the base of the lunar crust. The two rocks consist mostly of impact melt rock, which we believe represents the solidified melt that formed by large-scale melting of the lunar crust during the Imbrium basin impact.

The bulk elemental composition of this melt rock is very different than the average composition of lunar surface material, as determined by analyses of lunar soil samples and by remote-sensing experiments. It is richer in iron and magnesium and poorer in calcium and aluminum. We suggest that this solidified melt formed by melting of rocks of the lower crust, and we infer that the composition of the lower crust is different than that of the upper crust.

These impact melt rocks contain a myriad of large and small unmelted mineral and rock fragments (Fig. 2), most of which seem to be pieces of shattered pristine igneous rocks. We have studied these mineral and rock fragments and have compared their elemental compositions to those of mineral grains in pristine igneous rocks. In contrast to our expectations, we find that most of these fragments come from a rather limited number of pristine rocks.

We also find that the bulk elemental composition of the parts of these rocks that represent solidified impact melt cannot be made by melting and mixing, in any proportion, the types of fragments included within the melt. This observation tells us that there are "missing" rock types — rock types that are present in the lunar crust but that are not present, in forms unaltered by impacts, in our collection of lunar samples.

Surprisingly, we also find that anorthosite (a rock that consists almost entirely of the calcium-aluminum silicate mineral anorthite) is virtually absent among the rock and mineral fragments. Most workers believe that anorthosite makes up at least half the volume of the upper part of the lunar crust. We feel that anorthosite originally made up the upper part of the crust in the Imbrium area. We hypothesize that anorthosite fragments were not incorporated in the Imbrium melt because this type of rock was removed by the effects of large impacts that took place in the area before the Imbrium impact.

The ages of lunar rocks may be determined by measuring the concentrations of naturally radioactive isotopes (which decay at known rates) and their daughter products, and comparing these concentrations to the concentrations of stable isotopes. A surprising result of some of the early determinations of lunar rock ages was that the ages of impact melt rocks and of fragments from highlands breccias tended to cluster around 3.8 to 3.9 billion years. This clustering of ages suggested to Gerald Wasserburg and his co-workers (California Institute of Technology) that the Moon underwent a massive impact bombardment, by many relatively large objects, 3.8 to 3.9 billion years ago. They termed this event "the lunar cataclysm."

Many others workers questioned this interpretation, because it is difficult to understand why such a great number of large objects should be present in the solar system so long after its formation (about 4.6 billion years ago), and why so many of them should hit the Moon within a relatively short time interval. These other workers proposed that basins were produced on the Moon from the time of its formation until about 3.8 billion years ago, but that the formation of the latest basins obscured the record of formation of earlier basins.

Recently, the cataclysm concept has been revived by Graham Ryder. Ryder notes that the ages determined on melt rocks that are thought to represent basin impact melts are all in the 3.8-3.9 billion year age range. He suggests that, if there were older melts produced by basin-forming impacts, we would see some record in the spectrum of measured ages.

As part of our study of early cratering history, Jeff Taylor and I have collaborated with Tim Swindle (University of Arizona) and his co-workers to try to determine the age of a basin that has not previously been dated — the Crisium basin, on the eastern edge of the Moon. We searched through the samples brought to Earth in 1972 by the Soviet probe Luna 20, looking for impact melt rocks that might represent melt generated by the Crisium basin-forming impact. We analyzed several samples and found that the Crisium basin probably formed within the time interval from 3.8 to 3.9 billion years ago; our best estimate for the age of Crisium is 3.895 ±0.017 billion years.

Our estimate for the age of the Crisium basin provides support for the cataclysm concept. We note, however, that all of the basins that have been dated are close to each other, because the Apollo and Luna landing sites encompassed a fairly small area of the lunar surface. To resolve the issue of a cataclysm, ages are needed for basin impact melt rocks from areas on the Moon distant from the Apollo and Luna sites.

DRY CARBONATE FORMATION ON MARS: A PLAUSIBLE SINK FOR AN EARLY DENSE CO_2 ATMOSPHERE?

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Many ideas about the early martian climate involve the concept of a carbon dioxide atmosphere with a pressure of more than 1 bar in order to explain the presence of observed surface features which seem to be due to liquid water existing in a dense atmosphere. One bar is about the atmospheric pressure on Earth now (only a very small fraction of which is CO_2). The present atmosphere of Mars is mostly CO_2 and has a pressure of only seven thousandths of a bar. Carbonate formation would lower the CO_2 pressure and occurs easily in wet environments. This has been invoked to explain the destruction of a dense, early atmosphere on Mars. However, since surficial liquid water would only exist (in sufficient quantities) as long as the carbon dioxide pressure was above a level of about one atmosphere, an additional mechanism to further reduce the pressure of the early atmosphere must be identified.

To get it to the present value of seven thousandths of a bar, we have evaluated the concept of dry carbonate formation (i.e., without the presence of a liquid phase of water). Further, we have determined the implications of such a process for the evolution of the atmospheric pressure over time, and are presently beginning experimental work to evaluate the process in the laboratory.

The hypothesis that Mars once had a CO_2 atmosphere with a pressure much greater than today's value is not universally accepted because it rests on a chain of reasoning constrained mainly by images from orbiting spacecraft, and not by field observations. Spacecraft observations of branching valley networks (Fig. 1), similar to water runoff channels on Earth, and previous high erosion rates inferred from the degradation of craters and other

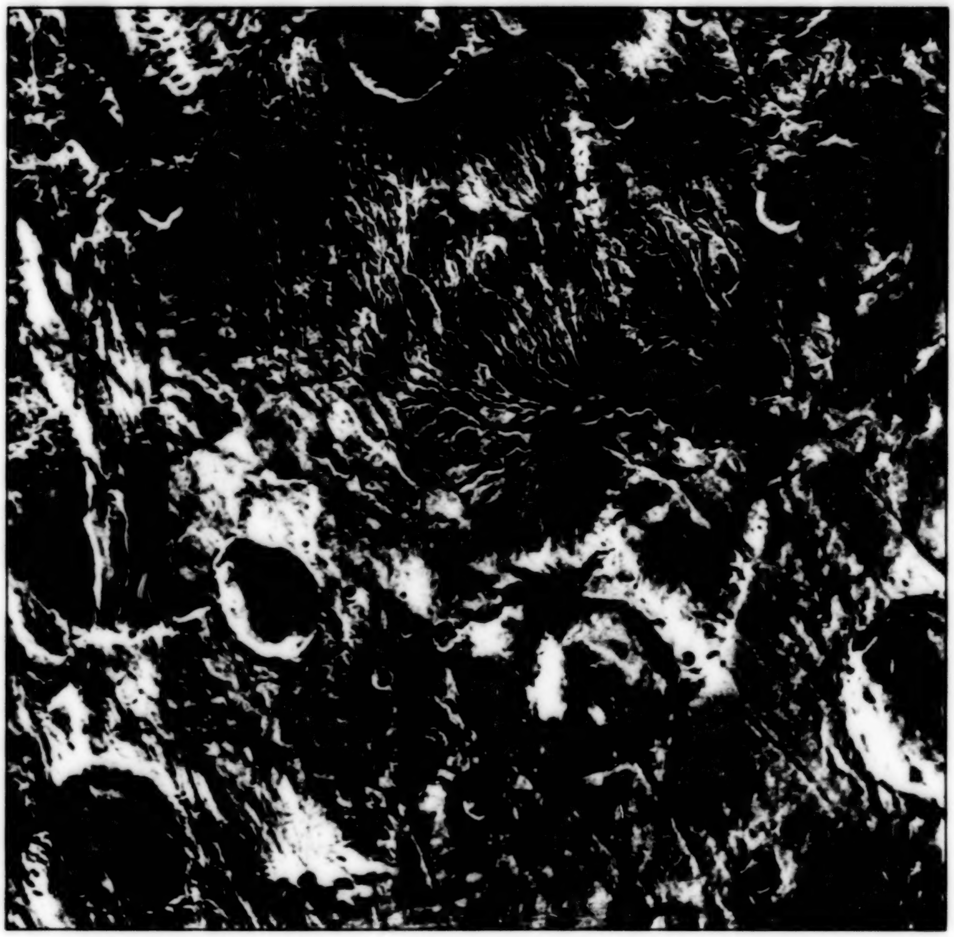

Fig. 1. Branching valley networks in the Southern Highlands of Mars, one piece of evidence for a possible early dense atmosphere which permitted liquid water to exist at the surface. (Viking Orbiter image. The largest crater is about 30 miles across.)

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features, both indicate the possible long-term presence on the surface of substantial liquid water (transitory rivers at least, lakes or oceans perhaps), although other explanations for some of the observations have also been proposed.

For liquid water to have existed at the surface over the millions of years thought to be necessary to explain the observations, the temperature most likely would have had to remain above the freezing point for that time. Given Mars' distance from the Sun, such a surface temperature could only have been maintained if the atmosphere was dense enough that greenhouse conditions existed. Carbon dioxide is a good greenhouse gas and an early dense atmosphere in contact with abundant surface water would have permitted the formation of carbonates in sufficient quantities to reduce the atmospheric CO_2 pressure.

However, there would have come a point as the temperature declined where the conditions would have been insufficient for further carbonate formation in this way. James Pollack (NASA Ames Research Center) and others have shown that the lowest pressure at which wet carbonate formation continues would have been around 1 bar (corresponding to a greenhouse temperature about equal to the freezing point of water).

The possibility that an early dense CO_2 atmosphere on Mars was destroyed by dry carbonate formation was first investigated in 1978 by Michael Booth and Hugh Kieffer (University of California, Los Angeles). At about the same time, work by James Gooding (University of New Mexico) showed that the reaction of CO_2 with silicate minerals to form carbonates was likely to be thermodynamically stable under past and present martian conditions — that is, it should proceed. However, the kinetics of the reaction — the *rate* at which it should proceed — was unknown. Booth and Kieffer performed an experiment to examine the rate of carbonate growth within rock powders subjected to Mars-like environmental simulations (including the near absence of liquid water).

Their resulting yield was less than a monolayer of carbonate on powder-size grains in the span of several days. However, they showed that the rate of the reaction more than accounts (by several orders of magnitude) for the destruction of 1 bar of CO_2 over geologic time (billions of years). Unfortunately, their study had major limitations; namely, there is no reliable basis for extrapolation to longer timescales, since less than a monolayer formed, and the nature of any "rind" (greater than one monolayer) which might have formed with further reaction was therefore not characterized.

Our treatment of the hypothesis of dry carbonate formation has been theoretical so far. We attempted to find published results where the carbonate formed by CO_2 interaction with silicates in the absence of liquid water was actually examined (e.g., with a scanning electron microscope), but without success. In the absence of experimental results, we considered the worst case — that in which the formation of a rind limits the effectiveness of the rapid reaction reported by Booth and Kieffer. Namely, we considered the question: Is the process (of dry carbonate formation) limited by the rate of reaction, or is it limited by the slow diffusion of CO_2 gas molecules through the carbonate rind which has formed? We found that the diffusivity of CO_2 through carbonate is probably low enough that the process is solidly in the diffusion-limited regime.

Over geologic time, this diffusion-limited dry carbonate formation is capable of forming coatings on silicate grains sufficient to store up to about 1 bar of CO_2, assuming reasonable values for the amount and size of available near-surface grains. Because we are dealing with a diffusion-limited reaction, Booth and Kieffer's dry carbonate growth rate could be too large by many orders of magnitude and this result would still hold. In this sense the reaction rate is irrelevant — it is the diffusivity that determines the rate of carbonate growth.

However, we would still like to verify the rapid reaction rate of Booth and

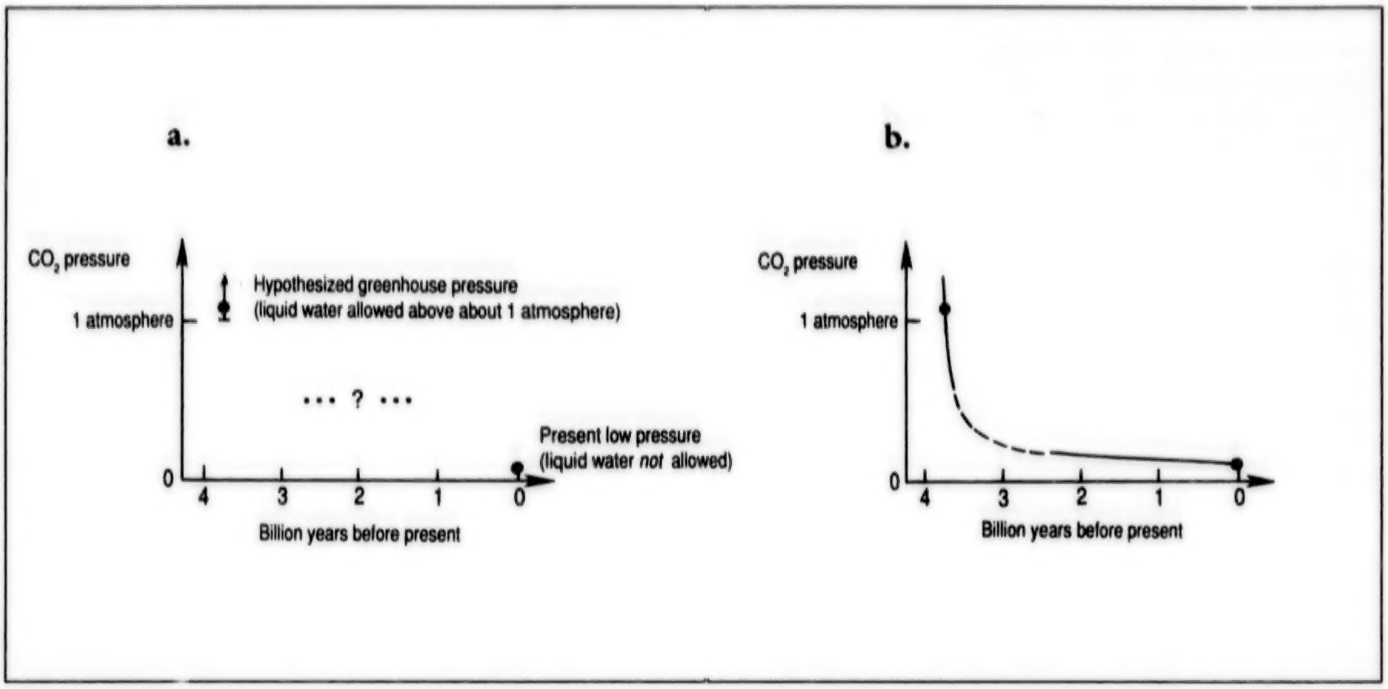

Fig. 2. a) The problem: If an early dense CO_2 atmosphere permitted carbonate formation in liquid water, then how did the pressure drop below 1 atmosphere to the present low pressure, where liquid water cannot exist? b) Dry carbonate formation: A possible solution? Note the predicted initial rapid drop, and the present-day slow decline in pressure.

Kieffer. We are presently proceeding with experiments which would improve on Booth and Kieffer's approach by (a) not constraining ourselves to present Mars surface conditions (e.g., using higher temperatures, hopefully expediting the reaction), and (b) better characterizing the reaction product (i.e., its composition, and whether the reaction indeed proceeds beyond a monolayer). Still, we have no guarantee that we will proceed beyond a monolayer, nor that Booth and Kieffer's results are a reliable guide for what we should expect.

In the meantime, assuming a diffusion-limited reaction, we have come to the following conclusions regarding the climate evolution of Mars (illustrated schematically in Fig. 2):

(1) Storage of CO_2 by this mechanism of dry carbonate formation could account for the 1-bar drop in pressure over geologic time.

(2) Because the diffusivity would have been higher when the pressure first began to drop (due to the higher temperature), the initial rate of CO_2 storage in carbonates would have been quite rapid.

(3) As CO_2 was stored and the pressure dropped, diffusion would have become more difficult, and the initial rapid rate of CO_2 pressure drop would have slowed drastically due to the strong influence of diffusivity on temperature (temperature follows pressure in a greenhouse scenario). Specifically, the present-day CO_2 pressure is predicted to be slowly decreasing (ignoring climate fluctuations and other sources of CO_2).

Further modeling of diffusion-limited dry carbonate formation, and a better understanding from our experiments of the rate of the reaction, will allow us to decide whether diffusion or the reaction rate determines weathering on Mars.

If the dry carbonate reaction is significant, it may help explain the transition from an early dense atmosphere to the present low pressure, and it may play a role in the present climate.

LUNAR METEORITES: EXPLORING THE MOON VIA ANTARCTICA

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The Moon is the only body in the solar system that can be positively identified as the source of a particular type of meteorite. Thanks to the sampling achieved by the U. S. Apollo program (manned landings at six separate sites between 1969 and 1972), and by the Soviet Luna program (automated core samplers at three sites between 1970 and 1976), we know that lunar materials have many distinctive properties which, observed in a meteorite, allow a definite identification of the Moon as the source. If we had not gone to the Moon, we would not be able to recognize our satellite as the source of these distinctive meteorites.

Rocks brought back from the Moon by the Apollo and Luna missions fall into two broad categories, on the basis of their mineral assemblages and geologic occurrences. The first category consists of rocks rich in the mineral plagioclase (calcium-aluminum silicate), collected from lunar highlands areas. The lunar highlands are the light-colored, densely cratered, relatively old areas of the lunar surface, and most highlands rocks show the effects of a long history of alteration by impacts. The second category consists of rocks rich in the mineral pyroxene (calcium-iron-magnesium silicate), collected from lunar maria areas. The lunar maria, or seas, are the dark-colored lava plains that cover the low-lying areas of the lunar surface. Their rocks, which are solidified lavas called mare basalts, are generally younger than highlands rocks.

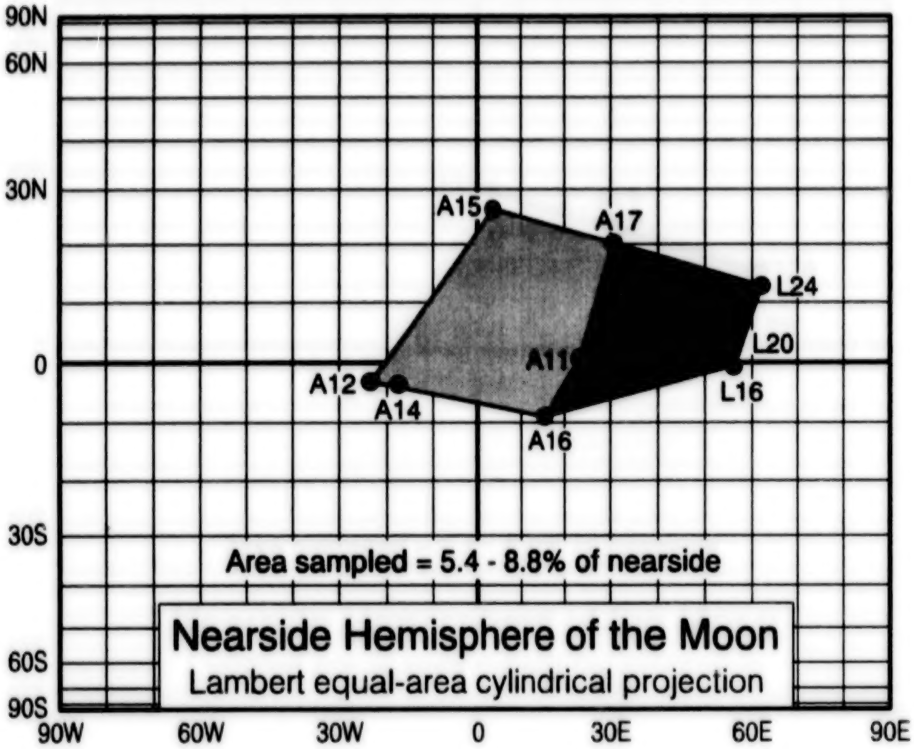
The Apollo and Luna missions also showed us that the material at the lunar surface is a kind of soil, called regolith, that consists of fragments of highlands and mare rocks broken, mixed, and melted by small impacts. Some of the lunar rocks brought to Earth are consolidated lunar soils, and these rocks are called regolith breccias.

The first lunar meteorite was recognized as such in 1982. Today, the number of lunar meteorites has climbed to eleven, and all of them have been found in Antarctica. However, the tally of truly separate lunar meteorites is reduced to eight, or perhaps fewer, if instances of "paired" meteorites (meteorites broken apart during passage through the atmosphere or upon impact with the solid Earth) are counted as single samples.

The Japanese collected five of the eight lunar meteorites, and they are stored at the Japanese National Institute of Polar Research (Tokyo, Japan). The other three, including the largest, were collected by the U.S. National Science Foundation and are curated at NASA's Johnson Space Center (Houston, Texas). The range of masses is typical for Antarctic meteorites, with five of the eight weighing only 2/10 of an ounce to 2 ounces. However, the biggest three weigh from 1 to 1.6 pounds each. As only a fraction of an ounce is needed to accomplish a wide range of scientific studies, these rocks provide ample material for many investigations.

The collections of meteorites that are brought back from Antarctica nearly every year consist of hundreds of specimens. These rocks must be classified based on their superficial characteristics in order to identify the most interesting samples for detailed study. About 1 out of 1000 meteorites found in Antarctica has turned out to be from the Moon. A lunar meteorite is typically first recognized because the compositions, shapes, and relations of its minerals are like those of the Apollo and Luna samples. Once a meteorite is identified as a potential lunar sample, then it is studied in detail. Various elemental ratios have been found to distinguish lunar rocks from other rocks. Particularly useful are the ratios of iron to manganese, gallium to aluminum, sodium to calcium, phosphorus to lanthanum, and cobalt to chromium. Even more definite signs of lunar affinity can be obtained by measuring certain isotopic ratios, most notably ratios among the three stable isotopes of oxygen (oxygen-16, oxygen-17 and oxygen-18).

The lunar meteorites that were first recognized as such are regolith breccias (consolidated lunar soils) that are rich in highland rock fragments. Because these meteorites are so rich in the mineral plagioclase, the possibility that they were lunar rocks was recognized fairly quickly after they were collected. During the last


90N
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A15
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L24
L20
A11
A12
A14
L16
A16
Area sampled = 5.4 - 8.8% of nearside
Nearside Hemisphere of the Moon
Lambert equal-area cylindrical projection

Fig. 1. The nine sites where Apollo and Luna samples were gathered are tightly clustered within the central portion of the lunar nearside (the side of the Moon that always faces the Earth). Thus, most of the lunar meteorites should be from areas not previously sampled.

year, four Antarctic meteorites that are basalts were also recognized to be lunar. Lunar mare basalts are more difficult to distinguish from other types of meteorites than are lunar highland rocks. The rocks are superficially similar to a class of meteorites called eucrites. The first lunar mare-basalt meteorite recognized as such was classified as a eucrite, based on its superficial characteristics, until it was studied in detail in 1989. Shortly after the initial discovery that a lunar basalt had been found among the Antarctic meteorites, Keizo Yanai (National Institute of Polar Research, Tokyo, Japan) identified two additional meteorites that are lunar mare basalts.

Lunar meteorites are transported to Earth as fragments ejected from impact craters produced when an asteroid or comet collides with the Moon. A typical impact that brings a lunar rock to Earth may involve an impacting body roughly half a mile in diameter smashing into the Moon at an interplanetary velocity of 20,000 miles per hour, forming a crater some six miles in diameter. Only a small proportion of the total debris ejected from such a crater would be propelled toward the Earth. Fortunately for the inhabitants of Earth, collisions of this magnitude are rare (perhaps one per million years for the Moon, and two per million years for the Earth).

Cosmic-ray exposure-age data give us information on the number of impacts that have produced the lunar meteorites, on the amount of time it took for a given lunar meteorite to reach the Earth, and on the history of breakup of the meteorite during its trip to Earth. Most (though not all) lunar meteorites have much shorter cosmic-ray exposure ages than other meteorites. This finding indicates that, after they were ejected from the lunar surface, they were not exposed to cosmic rays in space for very long. A related type of measurement, called the "terrestrial age", dates the time since the meteorite arrived on the Earth's surface. The sum of these two ages gives the time since the meteorite left the Moon.

Such ages have so far been determined for only four of the eight lunar meteorites, by Otto Eugster (University of Bern, Switzerland) and Kuni Nishiizumi (University of California at San Diego). Their results show that one lunar meteorite (represented by a group of three "paired" meteorites) left the Moon about 11 million years ago, one left less than 200,000 years ago, one left roughly 170,000 years ago, and one left between 200,000 and 500,000 years ago. Thus, at least two and probably three or four source craters are represented by just these four meteorites.

The value of the lunar meteorite collection as an augmentation of previous sampling of the Moon can be appreciated by noting that the Moon has been directly sampled only at nine tightly clustered sites, six Apollo and three Luna sites. For technological reasons, the nine sites were all within a small region of the Moon's central nearside. A polyhedron drawn around all six Apollo sites covers only 2.7 percent of the Moon's surface, and adding the three Luna sites only stretches this coverage to 4.4 percent (Fig. 1). Furthermore, sampling at each one of the sites was limited to a relatively small area.

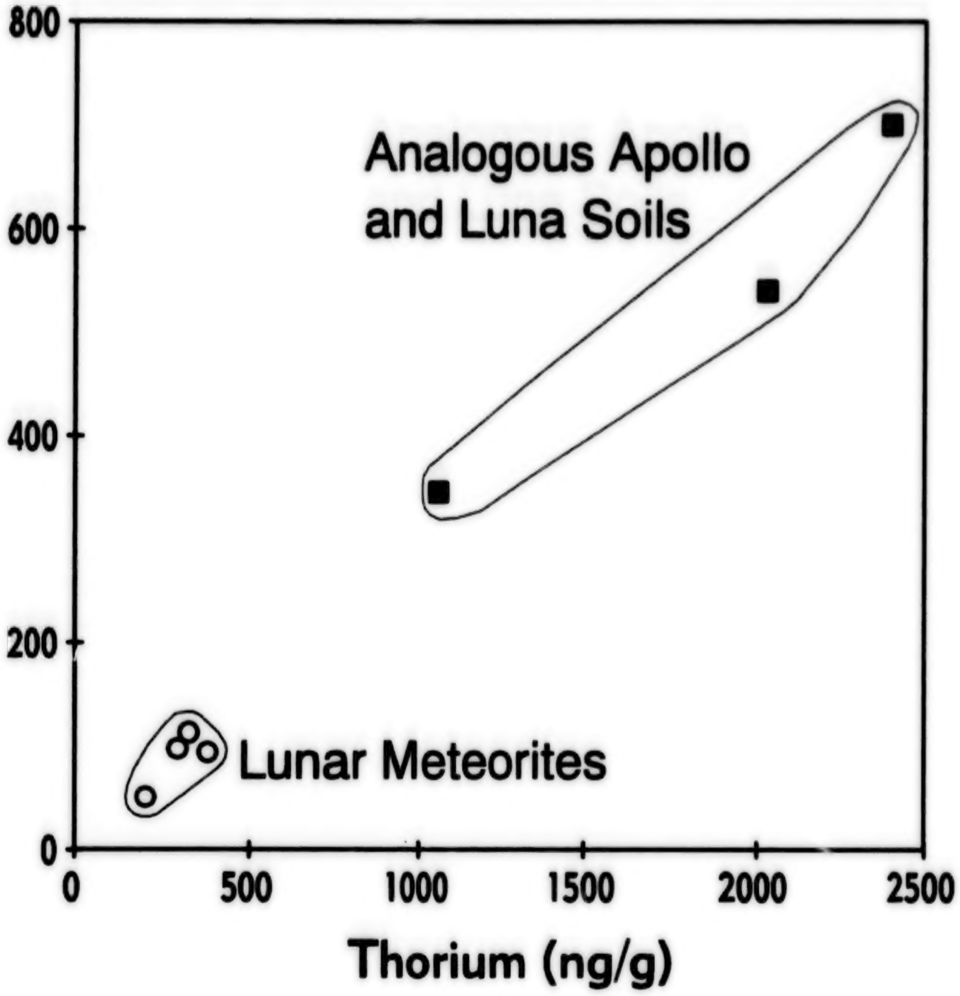

Fig. 2. Lunar meteorites have much lower concentrations of uranium and thorium than analogous samples from the Apollo and Luna landing sites.

During the later Apollo missions, the astronauts were able to obtain samples from as much as seven miles apart, but during the earlier missions all the samples were collected from very limited areas. The Luna probes acquired only one core from each landing site. Sampling of the Earth on a similar scale would yield only a small fraction of the rock types that occur on our planet.

Fortunately for lunar scientists, however, there is one aspect of lunar surface history that helps us obtain a more representative sample of lunar rocks than would be expected from the limited sampling area. Impacts on the lunar surface have broken rocks, ejected them from the areas where they formed, and strewn them over the lunar surface. Thus, a significant fraction of the rocks in the lunar regolith at any one place originally formed at points many tens, or even hundreds, of miles away. Even so, six mile-scale traverse zones plus three cores hardly constitutes adequate sampling of an obviously heterogeneous body with a circumference of 6800 miles.

Calculations indicate that the craters from which the lunar meteorites were ejected are probably distributed almost randomly across the lunar surface. Therefore, most of the lunar meteorites are probably not derived from the area we sampled with our lunar missions. The chances are good that at least one of the source craters is on the lunar farside, a part of the Moon where highlands predominate and maria are very rare.

The lunar meteorites show some distinctive compositional variations from the Apollo and Luna samples, which evidently are not very representative of the Moon as a whole. Some of these variations and their implications are as follows. First, the lunar meteorites that are regolith breccias are even more plagioclase rich than the average of Apollo and Luna highlands regolith samples. This finding provides additional support for the notion (popular since 1970) that the lunar crust formed when plagioclase floated to the top of a global "ocean" of magma on the primordial Moon.

Second, the lunar meteorites consistently have far lower contents of uranium, thorium and associated elements than most Apollo and Luna samples (Fig. 2). Apparently, the average lunar crust, and by extension the whole Moon, has considerably lower concentrations of these elements than previously assumed. Determination of the bulk-Moon content of these elements is important to evaluating models of lunar thermal history, because radioactive decay of uranium and thorium is the main long-term source of heat in a planet. It is also important in evaluating models of lunar origin, because some of these models predict large bulk-Moon enrichments for a whole class of elements that include uranium and thorium.

Third, the lunar meteorites also are consistently lower in siderophile elements (elements that tend to be associated with metallic iron) than similar Apollo and Luna samples. Several remarkable features of the ratios of siderophile elements were observed in the Apollo samples, for example, very high ratios of nickel to iridium and of gold to iridium. It now appears that these features are not bulk-Moon characteristics but are instead attributable mainly to unrepresentative sampling.

It is also remarkable that all of the mare-basalt meteorites are dissimilar in composition to most of the mare basalts collected by the Apollo and Luna missions. Most of the lunar mare basalts brought to Earth by these missions are significantly richer in titanium than the lunar mare basalts found as meteorites, and many are extremely rich in titanium. This finding suggests that the Apollo and Luna sampling of mare basalts was not representative.

A remarkably high proportion of the meteorites derived from the lunar highlands are consolidated lunar regolith. It seems that such rocks were greatly undersampled by the Apollo 16 mission (constituting only about 10 percent of the rocks brought to Earth). This was the one Apollo mission to a geochemically typical highlands site. In this sense, the lunar meteorites support previous speculations (for example by Paul Spudis, U. S. Geological Survey, Flagstaff) that the Apollo 16 sampling methods tended to avoid regolith breccias in favor of more cohesive crystalline rocks.

Within the highlands meteorites, there exist many small fragments of older rocks, a few of which represent highlands igneous rocks (rocks that crystallized from silicate melts, where the melts formed by internally generated heating of the planet). Systematic geochemical trends among highlands igneous rocks collected by the Apollo missions have been interpreted as evidence that igneous processes formed some of the rocks of the lunar crust long after the primordial magma ocean solidified. As we study the igneous rock fragments from the lunar meteorites and add the new information we obtain to the information we already have on the igneous history of the Moon, we look forward to a major test for current models of the evolution of the lunar crust.

COMETARY BOMBARDMENT OF THE EARTH

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The role of cometary impacts on the Earth has assumed new importance in the past decade. It was once thought that cometary nuclei were too small to contribute significantly to planetary cratering rates or to affect the volatile inventories of the terrestrial planets. However, the discovery in 1986 that the nucleus of Comet Halley was more than twice as large and ten times as massive as previously thought, has led to a new evaluation of the interaction of comets with the planetary system.

As a result, it is now recognized that comets likely contribute approximately one-third of the cratering in the terrestrial planets zone, the remainder coming from asteroids perturbed out of the main belt. The largest of these impacts may result in global biological extinction events which have the ability to change dramatically the evolution of life on the Earth. Perhaps most interesting, more than half of the cometary impacts come in brief but intense "cometary showers" during which cratering rates may rise by a factor of one hundred or more.

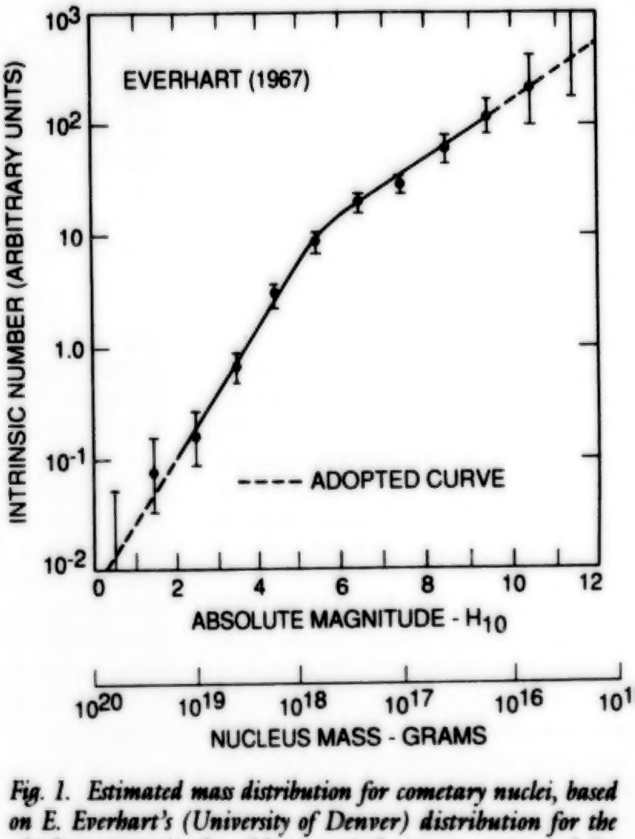

Fig. 1. Estimated mass distribution for cometary nuclei, based on E. Everhart's (University of Denver) distribution for the absolute magnitudes of long-period comets, corrected for observational selection effects.

Cometary cratering rates can be estimated by counting the number of observed comets crossing the Earth's orbit per unit time. The cratering is dominated by long-period comets (comets with periods >200 years) which are stored in distant orbits far from the planetary system in the Oort cloud. The Oort cloud contains on the order of 10^{13} comets larger than 0.7 mile radius. Approximately 10 of those comets cross the Earth's orbit per year, though not all are discovered, and there is an average probability of 2.2×10^{-9} that any random long-period comet will strike the Earth.

However, the cometary flux can vary widely over time. The strongest variations come from random passing stars actually penetrating the Oort cloud, severely perturbing the cometary orbits therein. In such instances, up to 7×10^8 long-period comets may be perturbed into Earth-crossing orbits in the space of less than 10^6 years. This intense shower of comets results in a sharp elevation in the total cratering rate. Fortunately, such intense shower events occur only about once every 5×10^8 years, though less intense cometary showers are expected more frequently.

Cratering rates can be estimated in the same manner for short-period comets (like Comets Halley and Encke) and for Earth-crossing asteroids. Results are shown in Table 1. About two-thirds of terrestrial cratering is by asteroids, and about one-third by comets.

These rates can be compared with the rate of cratering over the history of the solar system, estimated by counting the number of craters on planetary surfaces that have been dated using radio-isotope techniques. Currently, this is only possible for the Earth and the Moon. On the Earth, older craters have been removed by weathering and geophysical activity, and most recognized craters are less than 600 Myr old. From the number of craters, it is

Table 1. Cratering Rates on the Earth from Observed Fluxes

	Impacts per Myr*
Long-period comets	0.9
Short-period comets	0.4
Random comet showers	1.9
Earth-crossing asteroids (includes extinct cometary nuclei)	6.6
Total	9.8
Estimated from counted craters	
on the Earth (past 600 Myr):	11.2 ± 5.6
on the Moon (past 3.3 Gyr):	5.6 ± 2.5
Estimated cratering rate from major cometary showers every 26 Myr, plus the steady-state total.	30.6

*Impacts able to create craters ≥10km diameter.

estimated that approximately 11 craters larger than 6–mile diameter are formed every million years, though more than two thirds of the impacts are in the oceans and difficult to detect. This figure agrees well with the estimate for the total cometary and asteroidal cratering rate, shown in Table 1, though the possible errors are large.

The lunar maria preserve a much longer record, extending back over 3.3 billion years or more. The lunar cratering rate is roughly half that for the Earth, even after accounting for the Moon's smaller size and lower gravity. This suggests that the flux of impactors has recently increased. However, no mechanism for causing substantial long-term variations in the flux of asteroidal impactors is known.

There is a possible mechanism for varying cometary impactor rates. The perturbations on the Oort cloud which feed long-period comets into the planetary region result from random encounters with nearby stars and interstellar clouds. The Sun's current motion in the galaxy is slow when compared with other stars of similar mass and age. This lower velocity enhances the perturbations the cometary orbits receive from the passing stars. If the Sun had moved faster in the past, like typical G-type stars, then the perturbations would have been less and the resulting flux of comets into the planetary region would also have been less. In addition, if the Sun moved faster in the past, it likely spent more time out of the galactic plane, resulting in fewer stellar encounters overall.

Comets may also have played a role in the origin of life on Earth. Bombardment of the early Earth by comets during the formation of the solar system likely added a late veneer of volatiles with a total mass on the order of the current mass of the Earth's oceans and atmosphere. Along with these comets came complex hydrocarbons and other volatiles which were necessary for the creation of life. This "primordial soup" of pre-biotic materials provided the necessary chemical constituents for life to evolve, as soon as the early Earth had cooled to a sufficiently moderate temperature.

Some suggestions have been made that major impacts on the Earth and the resulting extinction events are periodic, occurring every 26 to 32 Myr. This suggestion is highly controversial but various hypotheses, each invoking cometary showers as the extinction mechanism, have been advanced to explain the alleged periodicity. These include: a faint, unseen solar companion star in a distant eccentric 26 Myr period orbit; a 10th planet at 100 to 150 AU from the Sun in an inclined, precessing orbit; or the Sun's epicyclic motion above and below the galactic plane.

Numerous dynamical problems exist with each hypothesis, rendering them all untenable. In addition, the expected cratering from the repeated cometary showers involved in these hypotheses would result in 5.5 times the observed record of lunar cratering over the past 3.3 Gyr. It is concluded that there is at present no workable mechanism for causing periodic impact driven extinctions every 26 to 32 Myr.

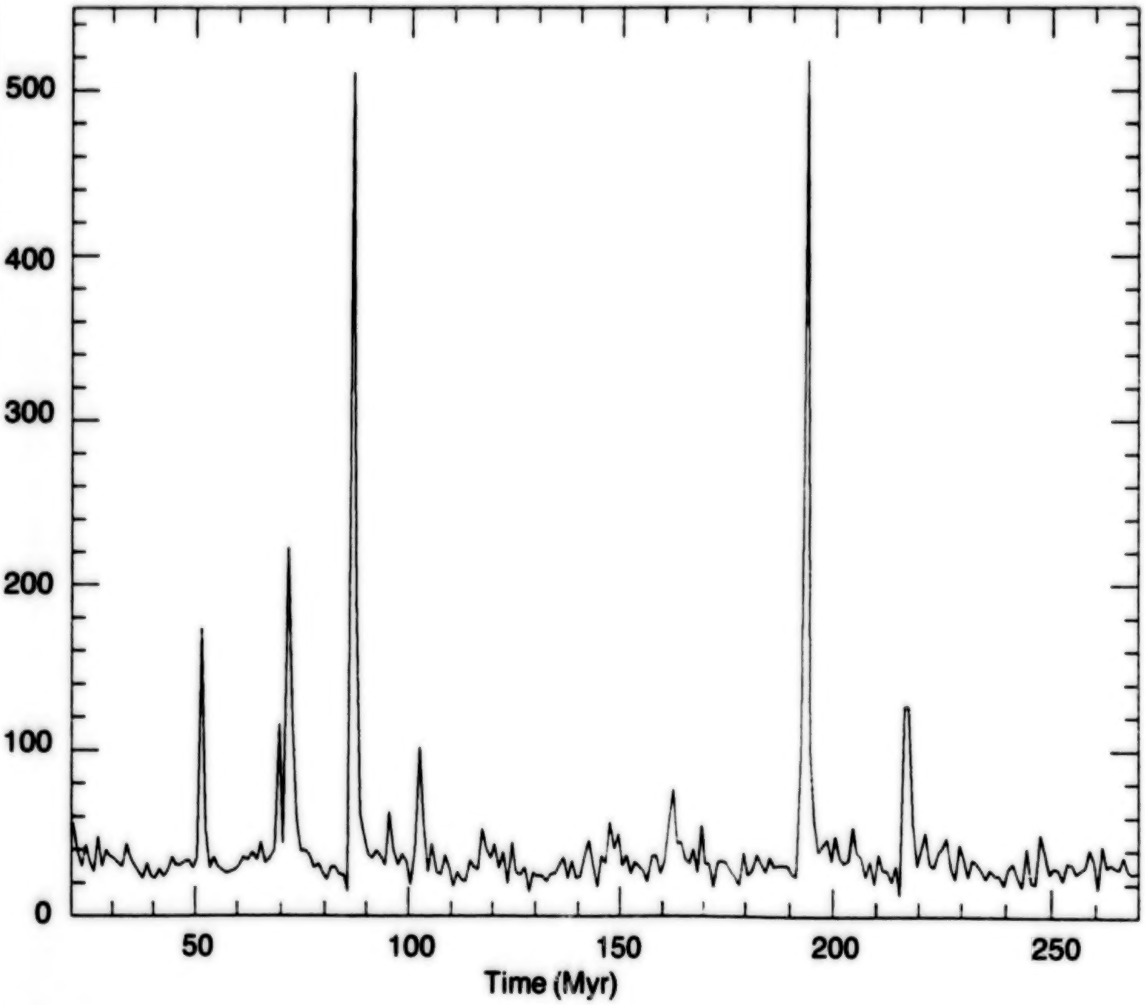

Fig. 2. Number of hypothetical comets passing within 2 astronomical units of the Sun based on a simulation of Oort cloud dynamics by J. Heisler (University of Arizona). The sharp spikes are cometary showers resulting from random stars passing directly through the Oort cloud.

USE OF THE LDEF SATELLITE TO ILLUMINATE THE NATURE OF COSMIC DUST AND SPACE DEBRIS PARTICLES IN LOW EARTH ORBIT

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After nearly 6 years in space, the Long Duration Exposure Facility (LDEF) was returned to Earth by the Space Shuttle Columbia in January of 1990. The recovery of LDEF is shown in Figure 1. As originally envisioned, LDEF was intended to be a bus-sized collector for cosmic dust and space debris particulates. However, as the project matured, and as NASA grew to realize the potential hazard posed to spacecraft by space environmental factors, other critical science and engineering objectives were added.

By its launch date in 1984, the mandate of the LDEF mission was to sense and record the nature of the total environment in low-Earth orbit, where most space-faring operations occur. The effects of cosmic dust and space debris impacts, solar and galactic radiation, oxygen erosion (the destruction of materials through collisions with oxygen present in the Earth's upper atmosphere), the relative vacuum of space, and other space-related phenomena are recorded onto the more than 10,000 test specimens flown on LDEF. These test specimens are now being studied by more than 300 investigators representing 7 NASA centers, 21 universities, 33 private companies, 9 DOD laboratories and 8 foreign countries.

Since the LDEF was gravity stabilized, the same end of the 30 foot long cylinder always faced the Earth, the same side always faced in the direction of orbital travel, and the opposite side always faced in the backwards (or trailing) direction. Therefore, careful examination of dust impact features, which are observed in abundance all over the satellite, will enable us to decipher the composition, velocity, and travel direction of each impacting dust grain.

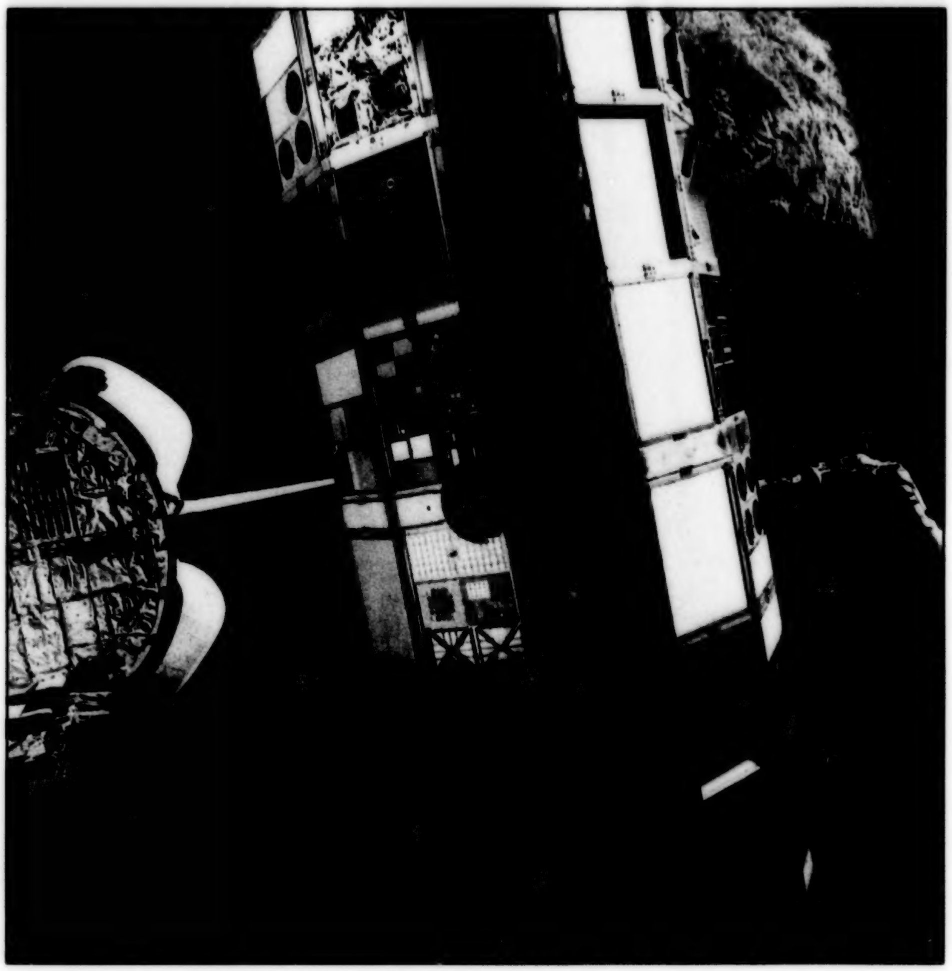

Fig. 1. The Atlantic Ocean serves as a backdrop for this Space Shuttle Columbia photograph of the recovery of the LDEF satellite, ending its 6 year flight in space. The recovered LDEF showed much damage attributable to the effects of dust impacts, radiation and oxygen erosion.

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This painstaking work will help to reveal, for the first time, potential sources of cometary and asteroidal dust particles. Such particles are predicted to retain the clearest record of the birth and early evolution of our solar system. In addition, the threat to spacecraft from these dust grains and particulate spacecraft debris can be evaluated. Therefore, analysis of the LDEF is expected to provide critical information for the design of future spacecraft as well as insight into Earth's origin.

Taking the first steps towards these goals, in the spring of 1990 a dedicated group of researchers labored to document all 5,000 of the largest dust impact features on the entire LDEF satellite. Figure 2 shows scientists engaged in this task. This work had to be completed before the satellite was dismantled, and scattered to laboratories all over the world. With this preliminary work successfully completed, coordinated detailed analyses have been initiated by space scientists in their home laboratories. These studies include compositional and isotopic analyses of dust remnants discovered within and about impact features (see Fig. 3), as well as the investigation of damage to space-exposed materials induced by impacting dust, solar and galactic radiation, and oxygen erosion.

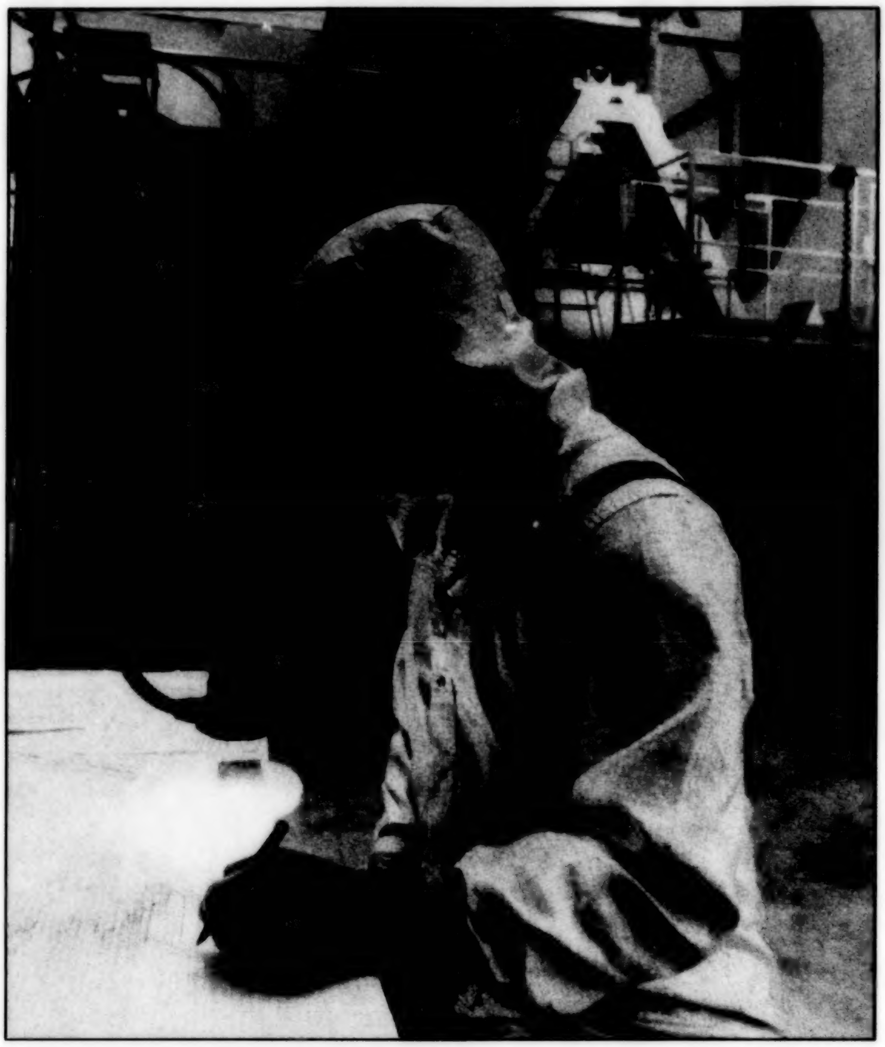

Fig. 2. In this view of a clean room at the Kennedy Space Center, a scientist is documenting large dust impact features on a portion of the LDEF satellite using a stereo microscope. The partially dismantled LDEF satellite is in the background.

To facilitate these and future analyses of LDEF test samples, many critical pieces of the LDEF have been curated at the Johnson Space Center. These space exposed surfaces take their place beside the lunar, Antarctic meteorite and stratospheric dust collections as a valuable resource for research on solar and planetary origins.

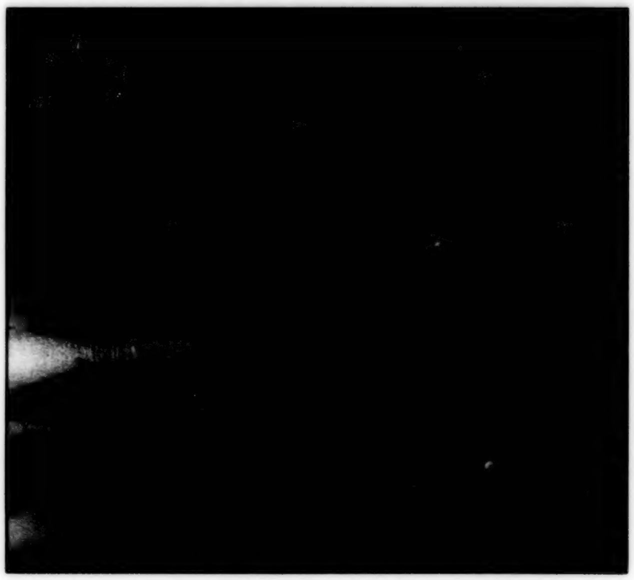

Fig. 3. This is a view of a large dust impact crater into an aluminum plate on LDEF. This crater measures approximately 0.5 inches long, and was probably made by a dust grain measuring no more than one-tenth this size. The red rays emanating from the crater are splashed remnants of the impacting dust particle, and their analysis will reveal the chemistry and source of this grain.

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